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Depositional and tectonic setting of the Archean Moodies Group, Barberton Greenstone Belt, South Africa

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Abstract

The 3.22–3.10 Ga old Moodies Group, uppermost unit of the Swaziland Supergroup in the Barberton Greenstone Belt (BGB), is the oldest exposed, well-preserved quartz-rich sedimentary sequence on earth. It is preserved in structurally separate blocks in a heavily deformed fold-and-thrust belt. North of the Inyoka Fault, Moodies strata reach up to 3700 m in thickness. Detailed mapping, correlation of measured sections, and systematic analysis of paleocurrents show that the lower Moodies Group north of the Inyoka Fault forms a deepening- and fining-upward sequence from a basal alluvial conglomerate through braided fluvial, tidal, and deltaic sandstones to offshore sandy shelf deposits. The basal conglomerate and overlying fluvial facies were derived from the north and include abundant detritus eroded from underlying Fig Tree Group dacitic volcanic rocks. Shoreline-parallel transport and extensive reworking dominate overlying deltaic, tidal, and marine facies. The lithologies and arrangement of Moodies Group facies, sandstone petrology, the unconformable relationship between Moodies strata and older deformed rocks, presence of at least one syndepositional normal fault, and presence of basaltic flow rocks and air-fall tuffs interbedded with the terrestrial strata collectively suggest that the lower Moodies Group was deposited in one or more intramontane basins in an extensional setting. Thinner Moodies sections south of the Inyoka Fault, generally less than 1000 m thick, may be correlative with the basal Moodies Group north of the Inyoka Fault and were probably deposited in separate basins.

A northerly derived, southward-thinning fan-delta conglomerate in the upper part of the Moodies Group in the central BGB overlies lower strata with an angular unconformity. This and associated upper Moodies conglomerates mark the beginning of basin shortening by south- to southeast-directed thrust faulting along the northern margin of the BGB and suggest that the upper Moodies Group was deposited in a foreland basin. Timing, orientation, and style of shortening suggest that this deformation eventually incorporated most of the BGB into a major fold-and-thrust belt.

1. Introduction

Archean greenstone belts preserve the world's oldest known well-preserved supracrustal rocks. Their stratigraphy, sedimentology, and deformational history have contributed significantly to our knowledge about the crustal history of the early earth. The BGB is one of the two best known pre-3.0 Ga old Archean greenstone belts (Lowe, 1982; Fig. 1).

The volcanic and sedimentary sequence comprising the Barberton Greenstone Belt, termed the Swaziland Supergroup (Anhaeusser, 1972),

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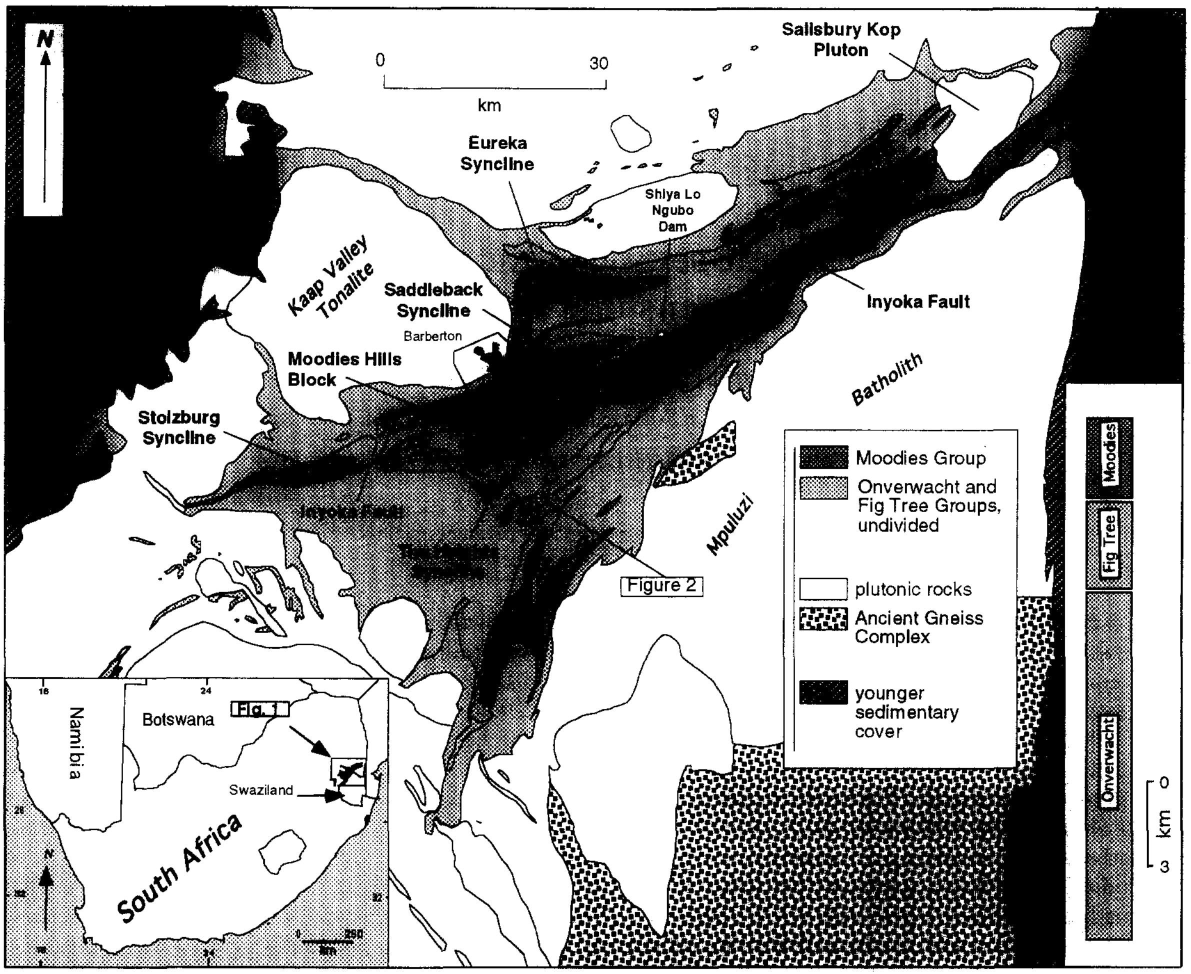


Fig. 1. Generalized geologic map of the Barberton Greenstone Belt (BGB) showing the location of the study area (Fig. 2). Inset to the lower right shows a highly generalized stratigraphic column of the BGB.

is subdivided, from base to top, into the Onverwacht, Fig Tree, and Moodies Groups (Fig. 1, inset). The first two form a structurally complex volcanic and immature, quartz-poor sedimentary sequence. The unconformably and/or structurally overlying Moodies Group consists of up to 3.7 km of alluvial to shallow-marine sandstone, subordinate conglomerate and shale, and minor volcanic units (Visser, 1956; Anhaeusser, 1976; Eriksson, 1978, 1980). Rocks of the Moodies Group form the world's oldest, relatively unmetamorphosed quartz-dominated sedimentary sequence, including quartz arenites, chert arenites, and arkosic sandstones deposited in alluvial to shallow-marine depositional settings. The quartz-rich composition and shallow-marine, deltaic, and subaerial depositional settings of the Moodies Group contrast strongly with the quartz-poor, immature, turbiditic greywackes of the underlying Fig Tree Group (Condie et al., 1970; Reimer, 1975; Eriksson, 1978).

Regional lithofacies and sedimentology of the Moodies Group have been described by Eriksson (1977, 1978, 1979, 1980) and Jackson et al. (1987). In this study, we will: (1) re-evaluate the stratigraphy and correlation of the structurally isolated blocks of Moodies strata in the central part of the BGB; (2) review the depositional setting of Moodies units based on new stratigraphic, paleocurrent, and petrographic data; and (3) reconsider the tectonic setting of the Moodies Group.

2. Geologic setting

2.1. Composition and age of the Swaziland Supergroup

Supracrustal rocks of the Swaziland Supergroup have been estimated to total 12 to 15 km in thickness (Viljoen and Viljoen, 1969; Anhaeusser, 1976), although some investigators have suggested that these estimates are inflated by tectonic repetition (de Wit, 1982; de Wit et al., 1983). Lowe and Byerly (1994) indicate an exposed thickness of 9 to 12 km for Onverwacht and Fig Tree rocks in the southern part of the BGB. The age of the sequence ranges from pre-3.5 Ga to ~3225 Ma for the Onverwacht and Fig Tree Groups and post-3225, pre-3110 Ma for the Moodies Group (Armstrong et al., 1990; Kröner et al., 1991; Kamo and Davis, 1994; de Ronde and de Wit, 1994).

The Onverwacht Group, 8 to 10 km thick, consists largely of komatiitic and tholeiitic volcanic rocks, minor felsic volcanic and volcaniclastic units, and thin interflow chert layers. The abundance of mafic and ultramafic volcanic rocks and the paucity of detrital quartz, high-grade metamorphic detritus, and continent-sourced sedimentary rocks have been interpreted to indicate deposition on a low-relief, low-energy, sediment-starved oceanic volcanic platform (Lowe and Knauth, 1977; Lowe, 1980, 1982). The Fig Tree Group, ~1–3 km thick, overlies Onverwacht Group rocks with sharp but apparently conformable contact. It consists largely of quartz-poor immature sandstones, tuffaceous and ferruginous fine-grained sedimentary rocks, chert, and banded-iron formation. U–Pb zircon ages from felsic volcaniclastic units indicate crystallization between ~3260 and 3225 Ma (Armstrong et al., 1990; Kröner et al., 1991; Kamo and Davis, 1944). Microcline and granitic rock fragments are present in Fig Tree clastic sedimentary rocks north of the Inyoka Fault, indicating erosion of granodioritic to granitic source rocks (Condie et al., 1970).

The uppermost unit of the Swaziland Supergroup, the Moodies Group, comprises up to 3.7 km of interbedded, compositionally immature to mature, dominantly quartzose sandstone, polymict conglomerate, and subordinate siltstone, shale, and volcanic rocks. The main framework components of Moodies Group sandstone are monocrystalline quartz, chert fragments, quartz–sericite-mosaic grains probably representing altered volcanic rock fragments, K-feldspar including microcline, and plagioclase (Heubeck and Lowe, 1994a).

2.2. Structure of the central BGB

The structure of the central BGB is dominated by a series of parallel, northeast-striking, tight to isoclinal, mostly overturned, northwest-verging synclines of variable plunge, 2 to 20 km in length and 1 to 10 km in width (Fig. 1). The synclines are separated from each other by tight anticlines or subvertical fault zones which probably represent collapsed anticlinal fold hinges (Ramsay, 1963). Most rocks of the Moodies Group are exposed in the synclines whereas polyphase-deformed rocks of the Fig Tree and Onverwacht Groups take up anticlinoria of variable width and fault zones. Bedding planes throughout the belt dip steeply or are overturned. Tonalitic, trondhjemitic, and granodioritic (TTG) intrusions deform but rarely intrude the margins of the belt (Anhaeusser, 1984). Most BGB–plutonic rock contacts are faults. Moodies strata on the limbs of most large folds have undergone negligible strain, as indicated by apparently undeformed conglomerate clasts, the paucity of mica reorientation and foliation in thin section, and the absence of shear or strain indications on bedding or joint surfaces. Deformation due to plane and shear strain, however, is significant near the margin of the greenstone belt, in the hinge zones of folds, and near major faults (Gay, 1969; Anhaeusser, 1969a, b; Lamb, 1984; Heubeck and Lowe, 1994b).

Regional metamorphism in the central BGB

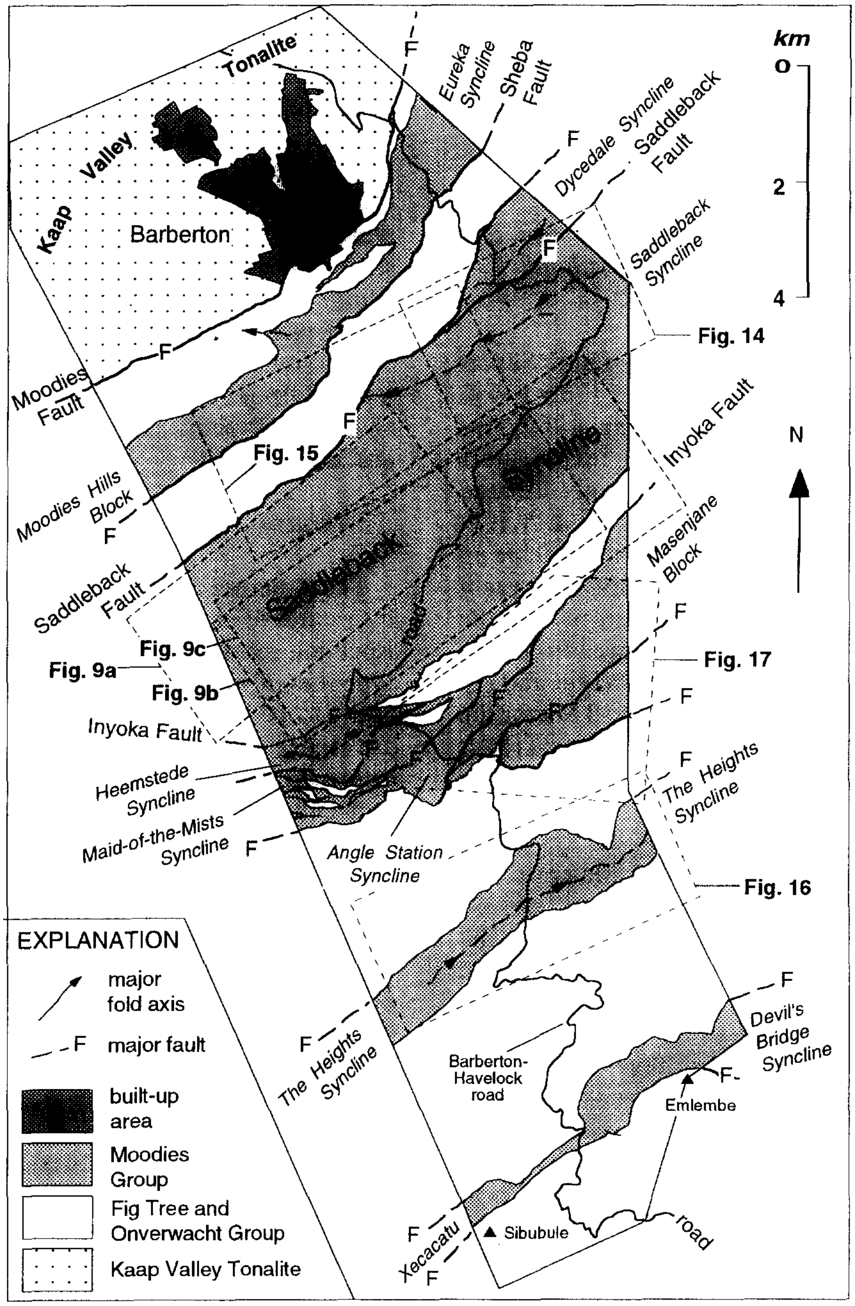


Fig. 2. Geologic map of the central BGB showing location of figures and geologic features.

reaches lower greenschist grade. Near the margins of the greenstone belt, contact metamorphism is locally important. In addition, rocks of the Moodies Group have been affected by post-depositional alteration including carbonatization, sericitization, and silicification.

The Inyoka Fault Zone, trending roughly axially through the greenstone belt, is the longest identifiable fault trace in the BGB and can be recognized continuously for more than 50 km from the western end of the belt, where the fault may continue westward into poorly exposed tonalitic plutons, through the central BGB, and into the incompletely mapped northeastern region (Fig. 1). Its significance as a stratigraphic divide between northern and southern Fig Tree facies was noted by Heinrichs and Reimer (1977), and the fault may have also separated tectonostratigraphic terranes during late evolution of the BGB (de Wit et al., 1992; de Ronde and de Wit, 1994; Heubeck and Lowe, 1994b). Juxtaposed rock units and truncation of fold limbs indicate that fault movement north of the Inyoka Fault was mostly up-to-the-north and that faults verged south (Daneel, 1987; Heubeck and Lowe, 1994b). Folds south of the Inyoka Fault face north (Heinrichs, 1980; Lamb, 1984; Paris, 1985). This geometry suggests that the Inyoka Fault may have acted as a major suture zone in the deformation history of the BGB (de Wit et al., 1992; de Ronde and de Wit, 1994).

The area mapped in detail for this study spans the central greenstone belt across the strike of the orogen between Barberton and the South Africa–Swaziland border, and includes major segments of most of the structural blocks exposing Moodies Group rocks (Figs. 1 and 2).

2.3. Age of the Moodies Group

Direct dating of the Moodies Group is difficult because it consists largely of clastic sedimentary rocks. A basaltic unit near the middle of the group is highly altered, and thin felsic tuffs in the lower part of the group have low zircon yields. Geochronologic dating of the Moodies Group in the past has focused on constraining its age relationships with datable older underlying or younger cross-cutting rocks.

The base of the Moodies Group unconformably overlies the top of the Fig Tree Group. The end of Fig Tree sedimentation is well constrained by dates from felsic and volcaniclastic units at the top of the group and from cross-cutting hypabyssal porphyritic rocks of Fig Tree age (Armstrong et al., 1990; Kröner et al., 1991; Kamo and Davis, 1994). Kröner et al. (1991) obtained an age of 3225 ± 3 Ma from fresh dacitic agglomerates underlying the base of the Moodies Group in the Stolzburg Syncline. Kamo and Davis (1994) report conventional zircon intercept ages of 3226 ± 1 and 3222^{+10}_{-4} Ma from an ignimbrite and porphyritic intrusion, respectively, at the top of the Fig Tree Group. These ages place the end of Fig Tree sedimentation between ~ 3228 and ~ 3222 Ma.

The onset of Moodies deposition is constrained by two data sets: (1) zircons from a dacitic dike cross-cutting the base of the Moodies Group in The Heights Syncline which were dated at 3207 ± 2 Ma in age (A. Kröner, pers. commun., 1993); and (2) the youngest near-concordant age of granitic clasts in the basal Moodies conglomerate in the northern BGB, which yielded an age of 3224 ± 6 Ma (sample Md-9 of Tegtmeyer and Kröner, 1987). These two data sets indicate that Moodies deposition began after 3224 ± 6 Ma but before 3207 ± 2 Ma.

The end of Moodies deposition is less well constrained. The Salisbury Kop Pluton which clearly cross-cuts deformed Moodies strata has been dated by Oosthuyzen (1970), Heubeck et al. (1993), and Kamo and Davis (1994) (Fig. 1). The most precise age of the crystallization of this pluton is 3109^{+10}_{-8} Ma (Kamo and Davis, 1994), indicating that Moodies deposition *and* deformation must have ended before that time. Ages obtained by Oosthuyzen (1970) and Heubeck et al. (1993) are less precise but consistent with the age of Kamo and Davis (1994).

In addition, two indirect data sets constrain the end of Moodies deposition. (1) Kamo and Davis (1994) obtained an age of 3216^{+2}_{-1} Ma for the Dalmein Pluton. This pluton cuts a large fold in Fig Tree and Onverwacht rocks which is similar in size, geometry, and orientation to folds formed during the major post-Moodies deformation. The age of the Dalmein Pluton may therefore possibly constrain the end of Moodies deposition. (2) Layer (1986) reported paleomagnetic data showing the basal Moodies Group to be magnetically overprinted by the Kaap Valley Pluton, dated at 3214 ± 4 Ma (Ar–Ar hornblende; Layer et al., 1992), suggesting that the Moodies Group was deposited and deformed by that time.

The available geochronologic data suggest that: (1) Fig Tree volcanism ended at $\sim 3225 \pm 3$ Ma; (2) Moodies deposition began no earlier than 3224 ± 6 Ma but had started by 3207 ± 2 Ma; (3) Moodies deposition and deformation had ended by 3109^{+10}_{-8} Ma and probably as early as 3214 ± 4 Ma.

3. Methods

The present study is based on detailed mapping of the central BGB at a scale of 1:6600, section measurement through the Moodies Group, description of sedimentary structures and textures, and systematic paleocurrent measurements (Heubeck, 1993). These data are supported by studies of sandstone petrography (Heubeck and Lowe, 1994a) and BGB structure (Heubeck and Lowe, 1994b).

Paleocurrent data were derived from tabular–planar foresets and trough cross-stratification. Ripple crest and clast imbrication data were collected in a few cases. In order to increase the accuracy of the recorded transport direction, three foresets were measured in each bedset of planar foresets where possible. Up to five readings from limbs of trough cross-stratification sets were measured where trough cross-beds were not fully exposed in three dimensions, and the trough axis was subsequently determined using a stereonet. Strongly asymmetrical or single-limbed troughs were not measured. Although the orientation of individual planar foresets in braided fluvial environments may diverge widely from the flow direction of the currents that formed them (Williams, 1966; Smith, 1972; Cant and Walker, 1978; Miall, 1985), the vector mean of a sufficiently large population is believed to generally identify flow direction in the main channel complex (Smith, 1972), and the maximum dip direction of planar foresets or the trend of the trough axis, respectively, were recorded as the mean sediment transport direction. Care was taken not to omit cross-beds from outcrops because of apparent abnormal directions, to avoid bias due to outcrop face orientation (Foster, 1991; Fedo and Cooper, 1991), and to sample outcrops regardless of the number of well-exposed paleocurrent indicators to ensure adequate regional coverage. All readings were corrected for tectonic fold plunge and tilt of bedding using rotation parameters which were determined through mapping and are shown in Fig. 3. Rose diagrams using a radial–linear frequency scale were constructed using *Stereonet* 4.4 (Allmendinger, 1992).

Because the Moodies Group is multiply deformed, incompletely exposed, and only partially preserved, a number of factors introduce variability in the collection and processing of the data (Pettijohn et al., 1987). The resulting inter-

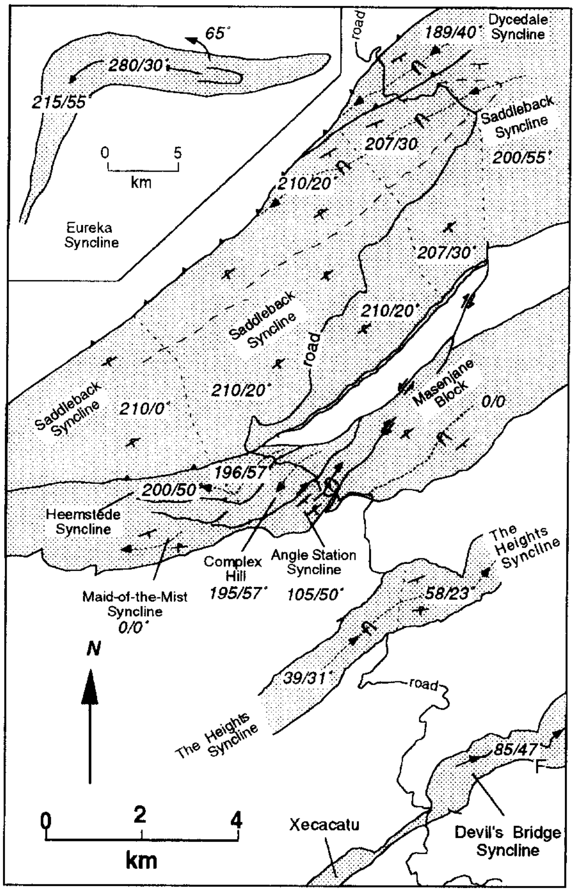


Fig. 3. Trend and plunge of regional fold axes in the Moodies Group in the central BGB used as rotation parameters to correct paleocurrent measurements. See Fig. 1 for location of Eureka Syncline.

pretation is therefore necessarily limited. Because some data sets are apparently not derived from an unimodal population and others may not conform to a circular-normal data distribution, we did not attempt to characterize sedimentary environments and dispersal patterns by methods of rigorous statistical analysis (Curray, 1956; Pettijohn et al., 1987; Le Roux, 1992).

4. Stratigraphy of the Moodies Group

4.1. Correlation

North of the Inyoka Fault, in the Eureka Syncline/Moodies Hills Block, Dycedale Syncline, Saddleback Syncline, and Stolzburg Syncline, the preserved Moodies Group includes between 0.5 and 3.7 km of strata (Fig. 4). Most are medium- and coarse-grained sandstone but conglomerate, siltstone, rare shale, and some banded-iron formation are present in most sections. A 0–20 m thick volcanic unit, the so-called Moodies Lava (MdL; Visser, 1956), is the only regionally identifiable time datum, and precise stratigraphic correlation above or below this unit is problematic (Fig. 4). However, regional stratigraphic changes in the relative abundance of K-feldspar, chert, monocrystalline quartz, and lithic fragments allow a broad petrographic correlation north of the Inyoka Fault (Heubeck and Lowe, 1994a).

Visser (1956), Anhaeusser (1976), and Eriksson (1980) have subdivided Moodies rocks north of the Inyoka Fault based on lithology, cyclicity, and/or sandstone petrography (Fig. 5). Initial regional mapping by Visser (1956) provided a lithologic breakdown of the Moodies Group. Anhaeusser (1976) refined the stratigraphy of Visser (1956) by emphasizing the cyclical or repetitive nature of Moodies strata. He defined three fining-upward lithostratigraphic cycles of formation status in the Eureka Syncline, each consisting of several mappable units (Fig. 5). Eriksson (1978) recognized five coarsening-upward cycles in most structural blocks north of the Inyoka Fault (Fig. 5). Three Moodies petrofacies were outlined by Heubeck and Lowe (1994a); sections north of the Inyoka Fault were subdivided into the Oosterbeek and the overlying Elephant's Head Petrofacies, and sections south of the Inyoka Fault were assigned to the Angle Station Petrofacies.

South of the Inyoka Fault, Moodies rocks are preserved in sections up to 0.6 km thick (Fig. 4), lack marker horizons, show only minor vertical petrographic changes, are mostly coarse-grained or gravelly, and lack K-feldspar. Plagioclase, which may have been present originally, has been completely altered to massive sericite and quartz–sericite-mosaic grains. Hose (1990) suggested that Moodies rocks south of the Inyoka Fault are correlative with the lower part of the Moodies Group north of the fault. While these sequences may be time-equivalents, based on the age of immediately underlying volcanic rocks of the Fig Tree Group, Heubeck and Lowe (1994a) have argued that the complete lack of detrital feldspar in Moodies Group rocks south of the Inyoka Fault suggests a source area lacking K-feldspar, in contrast to microcline-bearing rocks north of the fault. Deposition probably occurred in one or more basins that were isolated from that in which strata north of the fault accumulated. Because of the lack of unambiguous bases for correlation, sections south of the Inyoka Fault are not correlated in this study with those to the north or with each other (Figs. 2 and 5).

4.2. Moodies stratigraphy north of the Inyoka Fault

The standard section of the Moodies Group is located in the Eureka Syncline (Anhaeusser, 1976; Fig. 4) and includes ~3200 m of strata. The thickest known section is in the Saddleback Syncline where exposed Moodies rocks reach 3700 m thick (Fig. 4).

In the present study, Moodies units were defined and named following the subdivisions of Anhaeusser (1976). Eight main lithologic units are recognized in the Moodies Group north of the Inyoka Fault (Figs. 4 and 5). From base upward these are: (1) a basal polymict cobble conglomerate (MdB) 0 to 100 m thick, overlying sedimentary and volcanic rocks of the Fig Tree

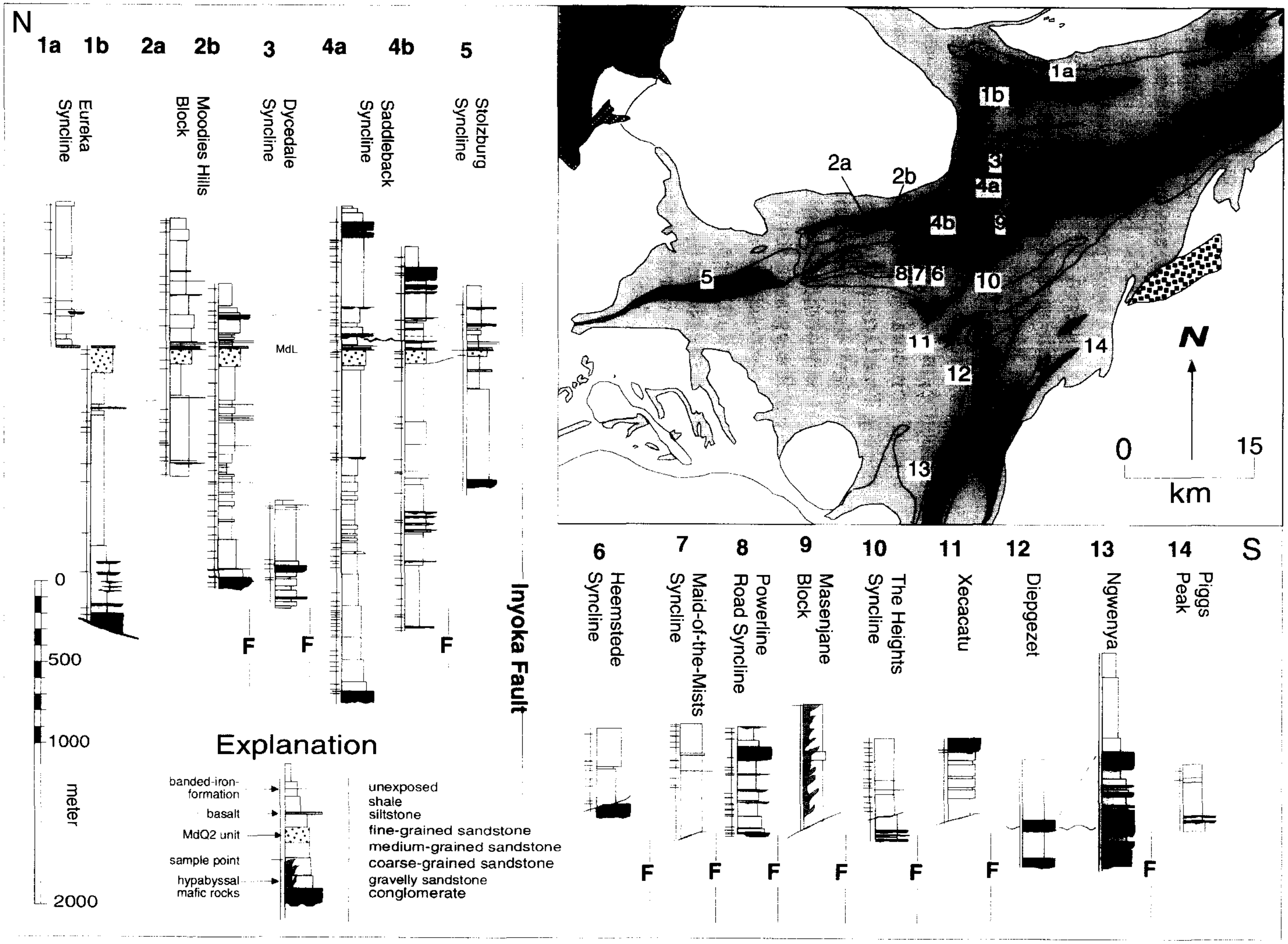


Fig. 4. Stratigraphic columns of the Moodies Group in the central BGB. Inset map shows approximate location of columns. Patterns in inset map are shown in Fig. 1. Columns are arranged N–S to allow for limited along-strike correlation. Columns 2a, 2b, and 4b were redrafted from Hose (1990), column 5 was redrafted from Reimer (1967), and columns 11–13 were redrafted from Lamb and Paris (1988).

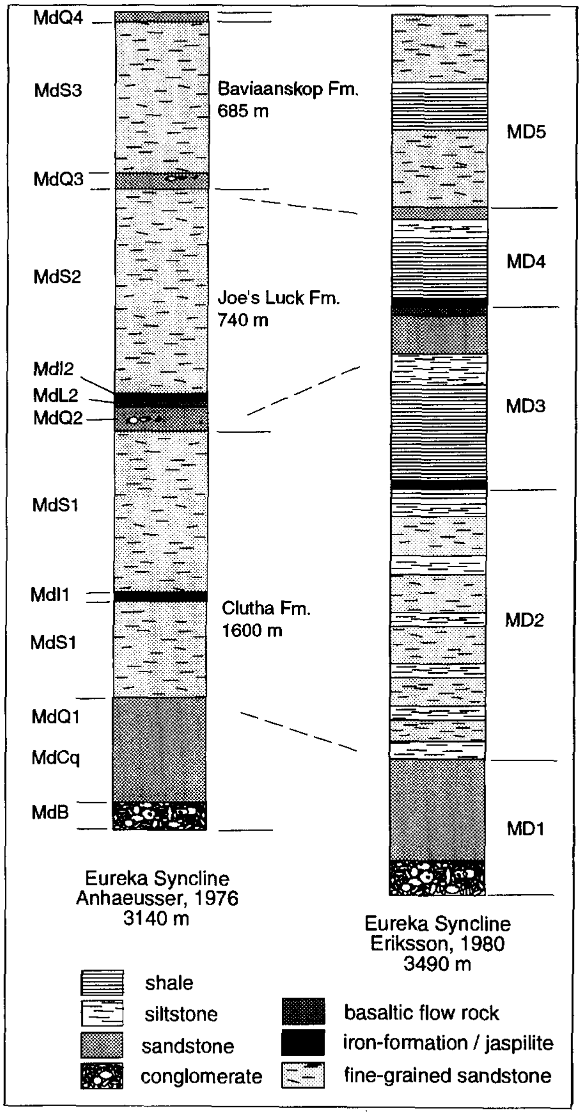


Fig. 5. Comparison of stratigraphic subdivisions proposed for the Moodies Group. Unit codes to the left follow the nomenclature of Anhaeusser (1976) and are adopted in this study.

and Onverwacht Groups; (2) 600–1100 m of coarse- and medium-grained, shallow-water, immature quartzose sandstone (MdCq and MdQ1); (3) 500–1000 m of generally poorly exposed fine-grained sandstone, subordinate siltstone, and rare jaspilitic and magnetic shale (MdS1 and MdI1); (4) coarse-grained quartz arenite (MdQ2) overlain by a thin jaspilite (MdI2); (5) metabasalt (MdL2); (6) a heterogeneous sequence of interbedded coarse-grained, shallow-water sandstone, tuffaceous siltstone, and minor conglomerate in the Saddleback Syncline, and shale with jaspilite, phyllite, and subgreywacke in the Eureka Syncline (MdS2); (7) cobble conglomerate in the Saddleback Syncline overlain by a fining-upward sequence, and conglomerate lenses and cross-bedded sandstone in the Eureka Syncline (MdQ3); and (8) a thick sequence of siltstone and shale (MdS3) capped by granule-bearing sandstone, preserved only in the Eureka Syncline (MdQ4).

Collectively, units (1) through (5) are termed the lower Moodies Group and units (6) through (8) the upper Moodies Group. The following discussion briefly describes the composition, sedimentary structures, distribution, and paleocurrent patterns of these units.

4.2.1. The basal conglomerate (MdB)

The Moodies basal conglomerate (MdB) consists of a clast-supported polymict pebble and cobble conglomerate interbedded with lenses of matrix-supported conglomerate, gravelly sandstone, and sandstone. It overlies tuff, tuffaceous sedimentary rock, volcanic rock, and feldspar porphyry of the Schoongezicht Formation of the Fig Tree Group disconformably or with a slightly angular unconformity. Its base is erosional, and local clast composition is related to the composition of immediately underlying rocks (Heubeck and Lowe, 1994a). The conglomerate forms a south- or southwestward-thinning wedge over an area larger than 15×30 km, grading into a gravelly sandstone in the western Saddleback Syncline (Fig. 6, Table 1). Further towards the southwest, a basal conglomerate is apparently absent. No paleocurrent measurements were obtained from the conglomerate. Both Anhaeusser (1976) and Reimer et al. (1985), however, suggest a northerly provenance at least for the basal conglomerate in the Eureka Syncline. Paleocurrents within overlying sandstones are south-directed (Fig. 7).

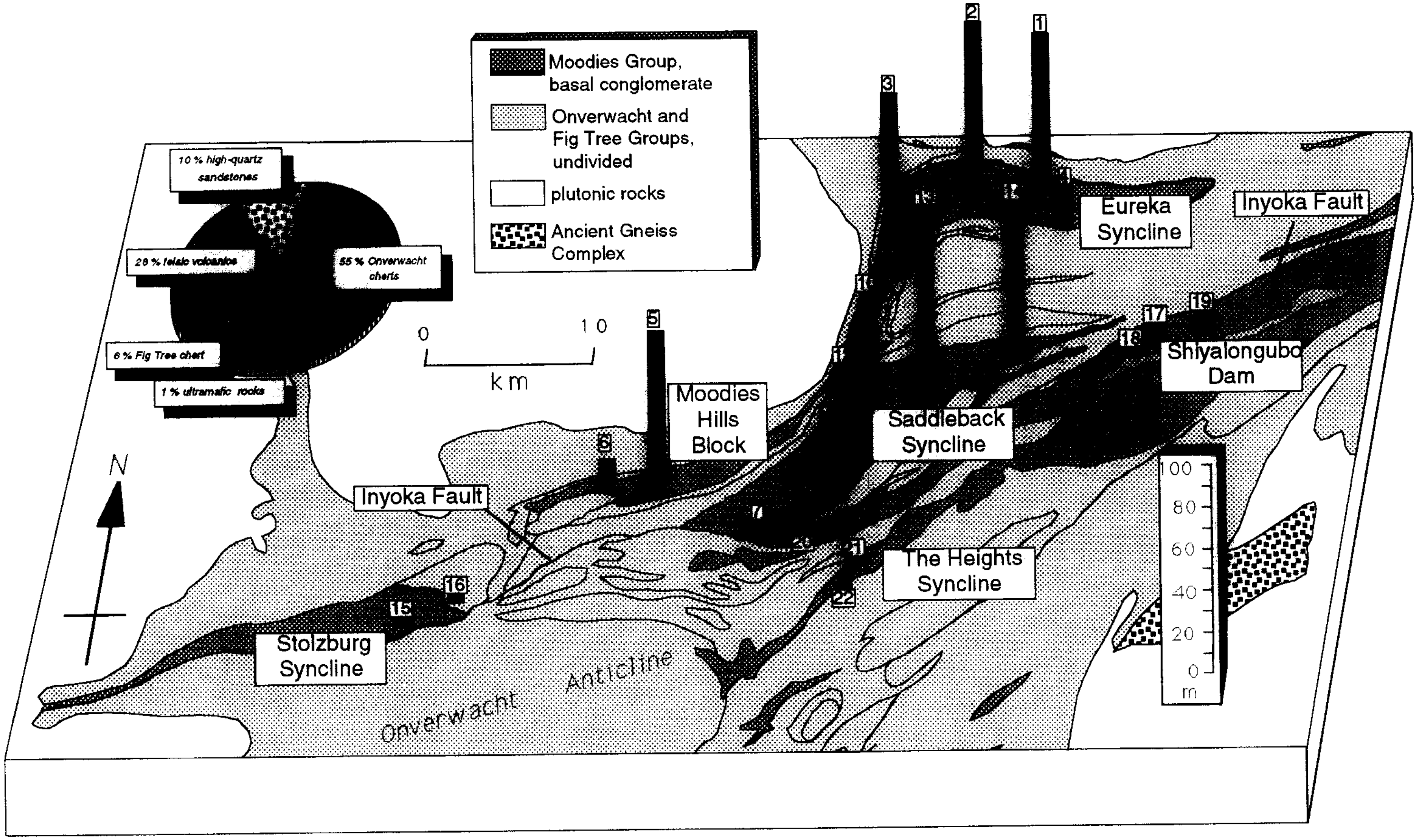


Fig. 6. Regional occurrence and thickness of the basal Moodies conglomerate in the central BGB. Inset pie diagram shows approximate mean clast composition of the Moodies basal conglomerate (from Heubeck and Lowe, 1994a). Granite clasts documented by Reimer et al. (1985) are common at sites 1, 2, and 3 but do not amount to more than 1% of overall clasts. Thickness trends suggest a southward or southwestward thinning of the conglomerate. Numbers on columns refer to locations listed in Table 1. The onset of Moodies sedimentation in the BGB may be diachronous, and the base of the conglomerate may not represent a time line (Lamb and Paris, 1988).

Table 1
Thickness of the Moodies basal conglomerate

Column	Location	Thickness (m)	Reference
1	Ezzie's Pass	70	Anhaeusser (1976)[a]
2	Joe's Luck Siding	70	Anhaeusser (1976)[a]
3	New Clutha	56	Anhaeusser (1976)[a]
4	Sheba Road	18	this study[a]
5	Agnes Mine	80	Hose (1990)
6	western Moodies Hills	20	D.R. Lowe (unpublished)
7	Skokohlwa	5	Hose (1990)
8	western Heemstede	7.5	this study
9	eastern Heemstede	40	this study
10/11	Oosterbeek	57–80	Visser et al. (1956); this study
12	Oosterbeek	60	Visser (1956)
13	Twello	98	Visser (1956)
14	Wonderscheur	74	Visser (1956)
15	Schoongezicht	5	Reimer (1967)
16	Belvue	5	Reimer (1967)
17	Shiyalongubo	5–10	this study
18	Shiyalongubo	4	this study
19	Shiyalongubo	3–18	Visser (1956)
20	Angle Station	2– 5	this study
21	The Heights Syncline	3– 6	this study
22	The Heights Syncline	4–10	this study

[a]Corrected for flattening strain.

The clast composition of the basal conglomerate changes rapidly along strike, even within the same block, indicating local provenance control (Eriksson, 1977; Reimer et al., 1985; Heubeck and Lowe, 1994a). The most abundant clasts are composed of resistant chert and other silica-rich rocks identical to those of the Onverwacht and Fig Tree Groups (Fig. 6). They also include variable proportions of felsic porphyries: silicified ultramafic rocks; lapilli tuff; Fig Tree tuffaceous shale, lithic sandstone, banded-iron formation and jaspilite; vein quartz; conglomerate; rare quartzose sandstone of unknown provenance; and rare granitic and metamorphic rocks. This last clast type has been reported chiefly from Ezzie's Pass in the northern Eureka Syncline (Visser, 1956; Reimer et al., 1985; Tegtmeyer and Kröner, 1987; Kröner and Compston, 1988). In contrast to the monotonous, nearly monomict Fig Tree chert-clast conglomerates, the Moodies basal conglomerate is distinctly polymict, nearly everywhere clast-supported, and has a sandstone matrix containing abundant monocrystalline quartz.

4.2.2. Lower Clutha Formation (MdCq and MdQ1)

The sandstone sequence gradationally overlying the basal conglomerate is known as the Moodies Quartzite (Visser, 1956). Anhaeusser (1976) distinguished a lower calcareous quartzite (MdCq) from an overlying feldspathic quartzite (MdQ1), although it is likely that much of the carbonate in the former is not an original framework component but was introduced during diagenesis or hydrothermal alteration (Anhaeusser, 1976). This sandstone sequence reaches its maximum thickness, ~1140 m, in the western Saddleback Syncline (Hose, 1990). The rock consists of thick- and medium-bedded, coarse- and medium-grained quartzose sandstone that is commonly granule- or gravel-bearing and interbedded with medium- to fine-grained sandstone. Conglomeratic layers are typically thin (< 1 m to single-clast thickness) and laterally discontinuous; they appear to be more common near the base of the unit. Individual sandstone beds can commonly be followed for tens of meters laterally; most end due to erosion

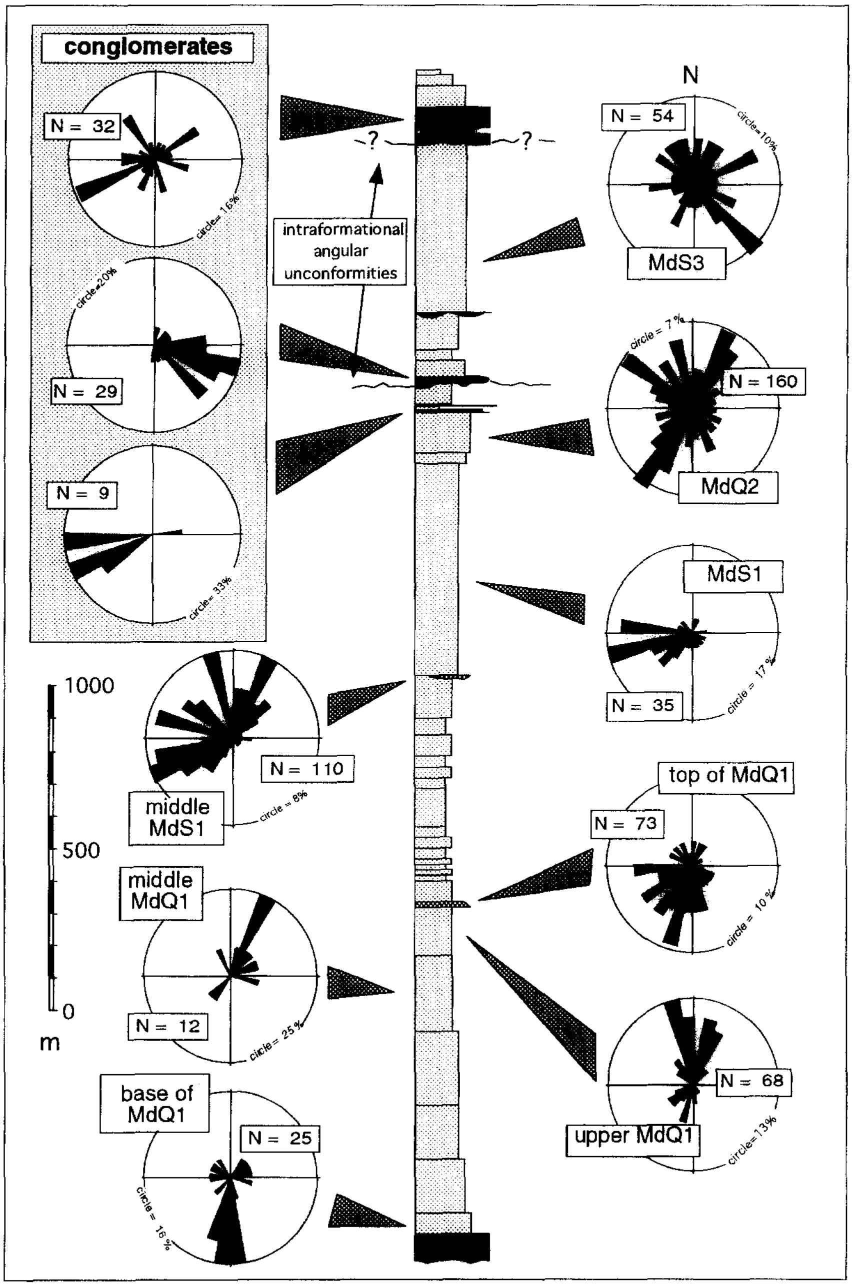


Fig. 7. Rose diagrams summarizing paleocurrent patterns of selected Moodies sandstone and conglomerate units discussed in the text, correlated with the Moodies stratigraphic column in the eastern Saddleback Syncline (column 4a of Fig. 4).

at the bases of overlying sandstone beds. Sedimentary structures include abundant low-angle planar and tangential foresets up to 1.2 m high, small-scale trough cross-stratification, planar bedding, clay drapes, and rare mudcracks. Where exposed, bedding planes commonly display straight-crested to sinuous ripples. Paleocurrents from the coarser basal part of this unit in the

Saddleback Syncline are mainly toward the south whereas paleocurrents higher in the section generally display a northerly mode (Fig. 7).

This sandstone sequence tends to become more quartzose upward. Relative abundances of matrix, chert, and unstable lithic grains decrease upward as the abundance of monocrystalline quartz increases from ~37% near the base to ~78% near the top. The proportion of feldspar, including plagioclase, microcline, and orthoclase, reaches a maximum of 20% toward the top of this unit (Heubeck and Lowe, 1994a).

The upper 250 m of MdQ1 are characterized by 2–10 m thick packages of thinly and undulatorily bedded sandstone interbedded with parallel-stratified and low-angle cross-stratified sandstone. The thinly bedded sandstone contains abundant crinkly, commonly stylolitized green and grey mud laminations spaced vertically from < 1 to 5–10 cm apart. The green and black mud laminations anastomose and bifurcate in cross-section (Fig. 8). Individual mud coats cannot be followed for more than 40 cm before they are truncated or merge with other mud layers. Scouring and erosional contacts are rare. In places, the mud coats are torn into wispy fragments or are individually folded or deformed into mushroom-shaped protrusions invariably directed stratigraphically up. Thin mud laminations also coat small-scale foresets. Fluid-escape structures up to 50 cm in height disrupt the sandstone and mud laminations. Isolated gravel stringers in the cross-stratified sandstone beds are of single-clast thickness and do not exceed 10 m in length. Paleocurrents from these horizons are bidirectional, with a northern transport direction dominating (Fig. 9a). Eriksson (1979) interpreted the interbedded sandstone–mud laminations as thin repetitive gravity flows on a steepened slope of a progradational fan. However, the anastomosing character of the mud laminations and association with cross-bedded sandstones appear to favor a shallow subtidal setting subjected to periodic changes in depositional energy. Moreover, a stratigraphically equivalent section in the Dycedale Syncline ex-

Fig. 8. Anastomosing thin mudstone layers separating fine- to medium-grained sandstones. This unit at the top of MdQ1 is interpreted as representing deposition in a shallow subtidal environment.

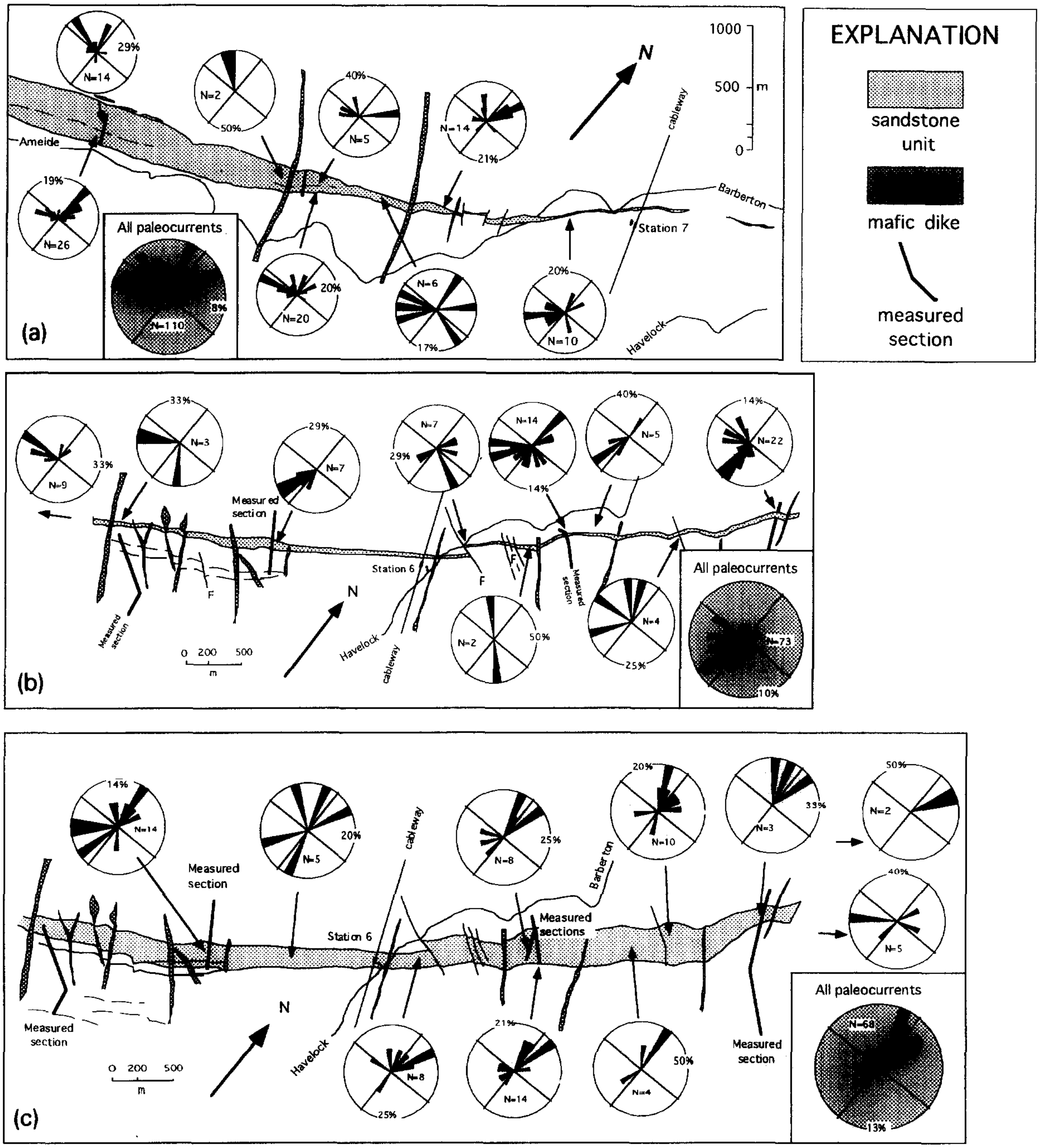


Fig. 9. Paleocurrent map of selected Moodies Group sandstones in the Saddleback Syncline (see Fig. 2 for location). All paleocurrents are corrected for fold plunge and tilt of bedding. (a) Tidal chert arenite in middle MdS1. (b) Large-scale cross-stratified, uppermost unit of MdQ1. (c) Tidal sandstone unit of upper MdQ1.

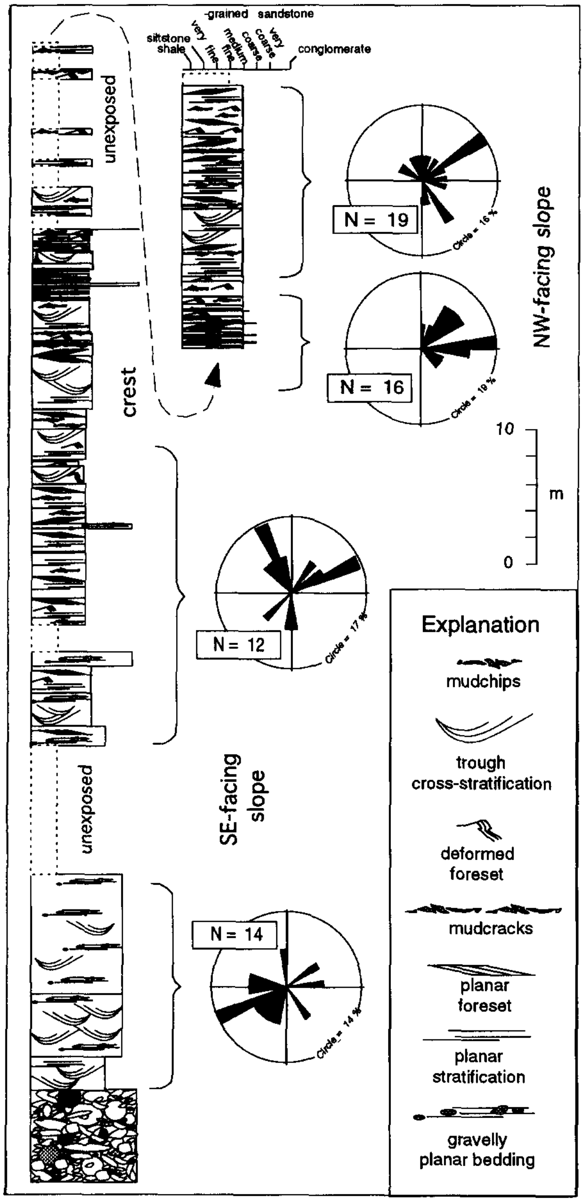


Fig. 10. Stratigraphic column and paleocurrent rose diagrams showing transition from fluvial to tidal facies near the top of MdQ1 in the Dycedale Syncline (Figs. 2 and 3).

poses planar and trough cross-bedded, fine- to coarse-grained sandstone which overlie mud-draped rippled bedding planes, many of which are mudcracked and can be followed for tens of meters (Fig. 10). Curled-up mudchips and angular granule-sized chert fragments are commonly incorporated in these trough cross-sets. The wide development of mud drapes in this facies suggests periodic quiescence and subsequent reactivation of bedforms, and small water-escape columns suggest rapid deposition and/or dewatering.

The uppermost unit of MdQ1 includes ~7 m of laterally continuous, medium-grained sandstone showing planar and tangential foresets reaching up to 2.4 m in height. The mean height of 215 measured foresets along a strike length of 7 km is 1.2 m. Paleocurrents in this unit are directed to the south (Fig. 9b).

4.2.3. MdS1

MdQ1 is overlain gradationally by a thick, fining-upward sequence composed mostly of medium- and fine-grained sandstone and siltstone with jaspilitic iron formation and magnetite-bearing shale members (Anhaeusser, 1976) and at least one air-fall tuff unit (C. Heubeck, unpubl. field data). The sequence reaches 790 m thick in the eastern Saddleback Syncline (Hose, 1990). It has been termed MdS1 by Visser (1956) and Anhaeusser (1976), the upper part of the Clutha Formation (Anhaeusser, 1976), MD3 (Eriksson, 1978), or Clutha B subdivision (Hose, 1990). In most areas, it is poorly exposed and deeply weathered. The paucity of exposure may be due to the abundance of easily weathered lithic detritus and the fine grain size of most sandstone in the unit. An air-laid felsic tuff is discontinuously preserved over a few tens of meters in the lower part of MdS1 in the Dycedale Syncline (Heubeck, 1993). A metabasaltic flow unit (MdL1) has been reported from the Stolzburg Syncline (Reimer, 1967), and reworked accretionary lapilli interbedded with sandy shales were found by Herget (1966) in the Moodies Hills Block. Bands of jaspilitic iron-formation interbedded with siltstone and magnetic shale and siltstones near the base of MdS1 have been reported from the Eureka Syncline (Anhaeusser, 1976), the Dycedale Syncline (Heubeck, 1993), the western Moodies Hills Block (Daneel, 1987; Hose, 1990), and the Stolzburg Syncline (Reimer, 1967). Similar units in the Saddleback Syncline may be covered by surficial deposits.

Outcrops of MdS1 are best and most numerous near its base. They show medium-grained sandstone with abundant planar stratification, ripples, and small-scale cross-lamination capped by thin mud layers. Paleocurrents from this unit in the Saddleback and Dycedale Synclines are largely directed to the southwest (Fig. 7). Columnar dewatering structures up to 2 m high are common in the basal part of this unit in the Saddleback Syncline and are also known from the Stolzburg Syncline (Reimer, 1967).

A chert arenite occurs near the middle of MdS1 in the Stolzburg Syncline (Reimer, 1967) and Saddleback Syncline, where it reaches ~60 m in thickness and wedges out towards the east. It is apparently absent from the Eureka Syncline. This "Lomati Quartzite" (Visser, 1956) is the type example of Eriksson's (1977) "medium- to coarse-grained sandstone facies". It is characterized by coarse- and very coarse-grained sandstone containing abundant small-scale bedforms including planar, trough, and climbing ripple cross-lamination, water-escape pillars, convoluted bedding, and rare mudchips. Paleocurrents are distinctly bimodal to the northwest and southeast (Fig. 9c).

4.2.4. MdQ2 and MdL2

The fine-grained sandstone and siltstone of MdS1 grade into a prominent, ridge-forming, medium- to coarse-grained quartz arenite (MdQ2) overlain by volcanic rock, the so-called "Moodies Lava" (MdL of Visser, 1956; MdL2 of Anhaeusser, 1976). Both units are present in the Stolzburg, Eureka, and Saddleback Synclines, and the Moodies Hills Block, over a distance of at least 60 km. They form the most important regional marker unit in the Moodies Group north of the Inyoka Fault.

The maximum thickness of MdQ2, up to 100 m in the Eureka Syncline, is likely to be in part structural (Anhaeusser, 1976). The typical thickness in the Saddleback Syncline is ~60 m. This unit is characterized by large, stacked planar and trough foresets near its top, with individual sets reaching up to 9 m thick (Fig. 11). In the Moodies Hills Block and the northeastern Saddleback Syncline, cross-stratification is subordinate to planar bedding (Daneel, 1987). Individual avalanche sets are 2–15 cm thick and normally graded. Bedding planes between foresets are commonly rippled but rarely draped by shale. Local conglomeratic lenses within MdQ2 occur in the Eureka Syncline (Anhaeusser, 1976). Paleocurrents from the Eureka and Saddleback Synclines show a regionally consistent southerly transport direction, although local flow toward the north occurs in the eastern Saddleback Syncline (Fig. 12). Paleocurrents from a discontinuous cobble conglomerate which overlies MdQ2 in the Saddleback Syncline also show southerly directions (Fig. 7).

MdQ2 or, where deposited, its capping conglomerate is generally conformably overlain by a metavolcanic basaltic unit (MdL2 of Anhaeusser, 1976; Daneel, 1987; Hose, 1990; G. Byerly, pers. commun., 1993). Reimer (1967) reports a thickness of 3–10 m of largely massive, vesicle-free metabasaltic flow rock overlain by a silty shale and a thin, ~0.3 m thick metabasalt in the Stolzburg Syncline. The same unit reaches ~8 m in the Moodies Hills Block and up to 22 m in the Saddleback Syncline, where Hose (1990) differentiated upper and lower flows based on variation in vesicle size and abundance. The rock is deeply altered to fine-grained chlorite, sericite, actinolite, carbonate, pyrite, and Fe-oxides. Vesicles are filled with quartz or carbonate. A thin volcanic tuff overlies the metabasalt in the Moodies Hill Block (Daneel, 1987), and jaspilitic bands (MdI2) overlie it in the Stolzburg Syncline, the Moodies Hills Block, the eastern Saddleback Syncline, and the Eureka Syncline. This distinctive volcanic unit, principal correlation unit for the Moodies blocks north of the Inyoka Fault, has not been identified south of this fault. A second, similar metabasaltic unit (MdL) occurs in MdS1 in the Stolzburg Syncline (Reimer, 1967; Fig. 4).

4.2.5. MdS2, MdQ3, MdS3, and MdS4

In the Eureka Syncline and the Moodies Hills Block, the stratigraphic sequence above the metabasalt is dissimilar to that in the Saddleback Syncline. In the Saddleback Syncline, the upper Moodies Group consists largely of sandstone with

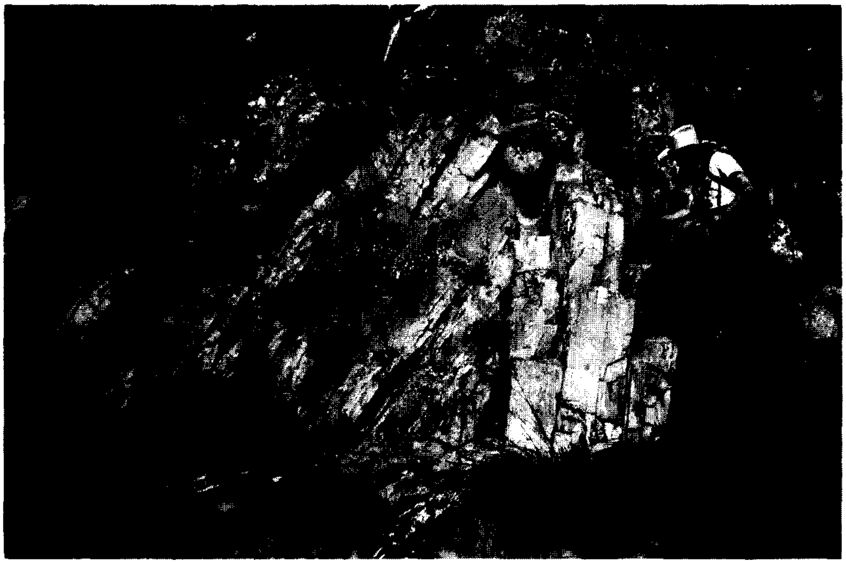


Fig. 11. Outcrop photograph of large-scale cross-bedded, coarse-grained sandstone of MdQ2 in the central BGB interpreted to represent part of a subaqueous dune field. Stratigraphic top is to the right. Saddleback Syncline.

some interbedded conglomerate. In the Eureka Syncline (Anhaeusser, 1976) and the Moodies Hills Block (Daneel, 1987), sandstone occurs mainly in a single, thick, cross-bedded and in places pebbly unit (MdQ3), with most of the upper Moodies Group, nearly 1000 m, composed of fine-grained sandstone, siltstone, and sparse shale (Anhaeusser, 1976; Eriksson, 1978; Hose, 1990).

A layer of air-fall tuff up to 2 m thick and lacking evidence of sedimentary reworking is preserved over a distance of a few hundred meters in the eastern Saddleback Syncline. Underlying sandstone also appears to be tuffaceous and is rich in fine-grained sericitic matrix.

Two conglomerate units, ~200 m and 600 m above MdL, respectively, occur in the Saddleback Syncline, the Stolzburg Syncline, and the Moodies Hills Block. The lower conglomerate overlies an intraformational angular unconformity in the Saddleback Syncline. Both conglomerates are clast-supported, although the upper conglomerate shows a rapid alternation of clast- and matrix-supported beds between 4 and 10 m thick (Eriksson, 1980). They are separated by several hundred meters of medium-grained, locally gravelly, quartzose sandstone with abundant planar and trough cross-stratification, mud coats, rippled bedding planes, and rare mudcracked zones. Paleocurrent patterns in these sandstones are highly variable but generally north-directed (Fig. 7). Above the upper conglomerate, rocks in the Saddleback Syncline fine rapidly upward into medium-grained sandstone. The highest preserved strata of the Moodies Group, exposed in the hinge of the Saddleback Syncline, consist of tuffaceous sandstone, magnetic shale, and minor BIF/jaspilite units. A jaspilite band associated with thick fine-grained sedimentary rocks also occurs in the Moodies Hills Block but only magnetic shale appears to exist in the Eureka Syncline (Anhaeusser, 1976).

In the Eureka Syncline, in contrast, the uppermost Moodies strata include ~1000 m of fine-grained shale, argillite, and litharenite capped by a locally developed, cross-bedded, slightly pebbly sandstone (MdQ4; Anhaeusser, 1976). Correlation of this thick fine-grained unit to the sand-dominated section in the Saddleback Syncline is problematic.

The stratigraphic top of the Moodies Group is nowhere preserved.

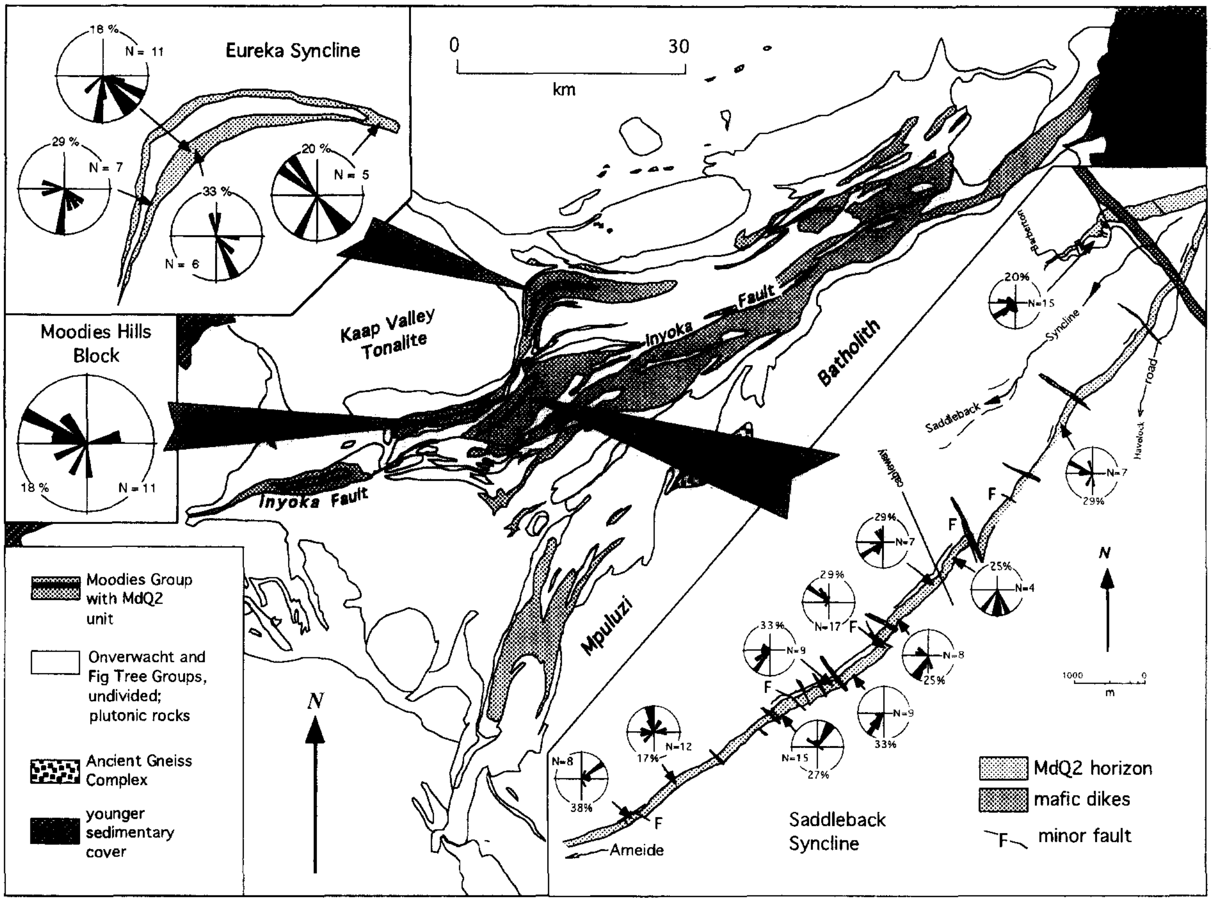


Fig. 12. Outcrop distribution and paleocurrent patterns of MdQ2 unit in the central BGB. Paleocurrents are predominantly toward the south and west, parallel to the present strike of the orogen, but show a strong, possibly local component to the north in the western Saddleback Syncline.

4.3. Intraformational unconformity at Saddleback Pass

An intraformational angular unconformity is exposed beneath a conglomerate in MdS2 in the upper Moodies Group in the Saddleback Syncline (Figs. 13 and 14). The angular relationship between bedding below the conglomerate and bedding within or immediately above the conglomerate decreases progressively to the east and south, away from the Saddleback Fault, and becomes conformable on the overturned limb of the Saddleback Syncline (Fig. 14). The conglomerate thins to the south away from the Saddleback Fault and grades into pebbly sandstone ~2 km from the fault. Paleocurrents from planar and trough foresets in thin lenses of gravelly sandstone within the conglomerate suggest a northern or western provenance. Paleocurrents in overlying cross-bedded sandstone are polymodal. Close to the Saddleback Fault, the base of the conglomerate truncates a highly wedge-shaped cobble conglomerate that is up to 14 m thick and crops out for only ~200 m (Figs. 13, 14). The composition of both conglomerates is identical and dominated by resistant rock types, including Fig Tree and Onverwacht chert and quartzose sandstone of Moodies affinity.

Similar features suggesting an unconformable relationship can be seen in a thick clast-sup-

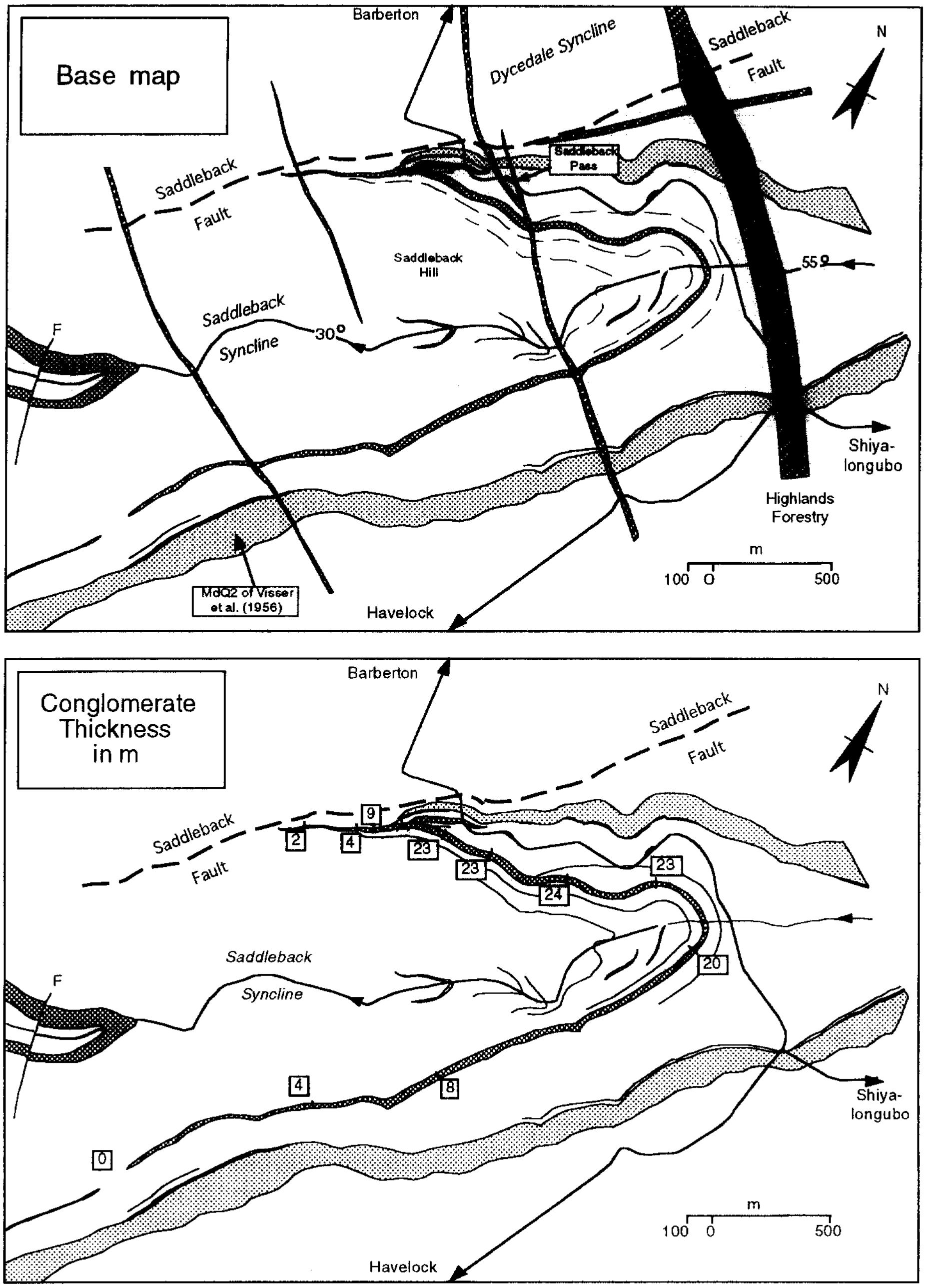
Fig. 13. Outcrop photograph of angular intraformational unconformity in the upper Moodies Group exposed at Saddleback Pass, Saddleback Syncline (Fig. 2). Strata dip moderately to steeply to the left (south). Tidal and deltaic sandstones (*a*) overlying an alluvial conglomerate (*b*) dipping steeply to the left at the right of the photograph are truncated and overlain by an alluvial conglomerate and overlying fluvial and tidal sandstones (*c*).

ported, chert-cobble conglomerate ~400 m higher in the section (Fig. 15). This conglomerate (MdQ3) reaches up to 200 m in thickness in the eastern Saddleback Syncline. When bedding in the now overturned conglomerate is restored to the horizontal, bedding in underlying rocks dips at a low angle towards the southeast (Fig. 15). Paleocurrents are widely divergent but generally south-directed. The conglomerate thins to the south, grading into discontinuous lenses of conglomerate and gravelly sandstone (Fig. 15). This conglomerate may be correlatable with a 5 m thick, matrix-supported conglomerate in the Moodies Hills Block but appears to be absent in the Eureka Syncline.

4.4. Stratigraphy of the Moodies Group south of the Inyoka Fault

Moodies rocks south of the Inyoka Fault are exposed in a number of small, probably once contiguous structural blocks along the Inyoka Fault, the The Heights Syncline, the Xecacatu Block, Devil's Bridge Syncline, and several areas in Swaziland, where they have been assigned to the Malolotsha Group (Lamb and Paris, 1988) pending correlation with the type Moodies rocks north of the Inyoka Fault (Figs. 1 and 2).

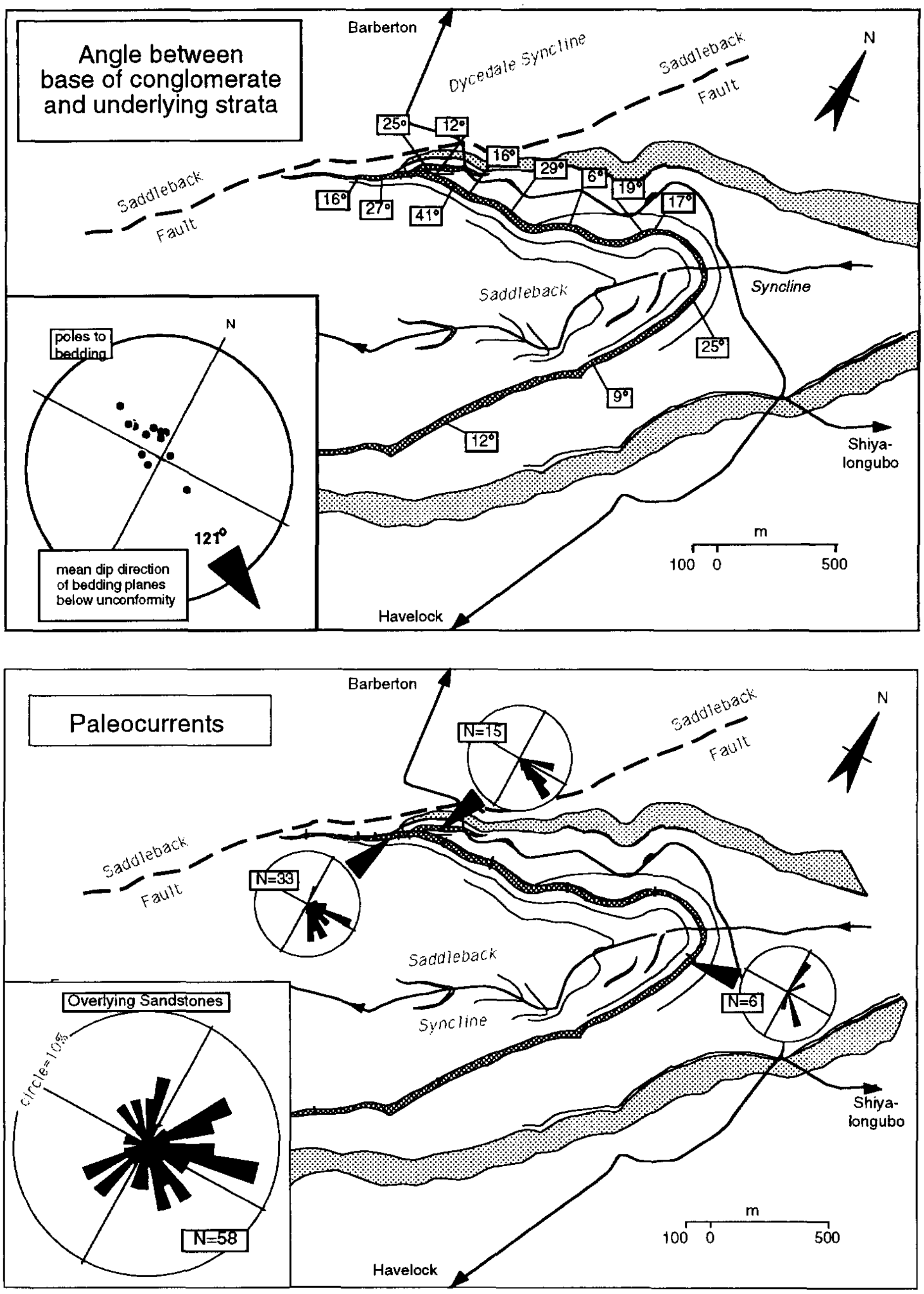
One of the largest outcrop belts of coarse clastic sedimentary rocks in the south-central BGB, the Sibubule–Emlembe belt that straddles the South African–Swaziland border, includes thick conglomerate units in the lower part of the Fig Tree Group that appear to represent alluvial fans (Fig. 2). The conglomerate, reaching up to 1 km in thickness, is a monomict chert-clast conglomerate, petrologically unlike Moodies conglomerate (Heubeck and Lowe, 1994a). It conformably overlies banded ferruginous chert, jasper, and ferruginous shale of the lower Fig Tree Group. Sandstone lenses in the conglomerate are petrologically identical to unambiguous Fig Tree chert arenites in Mapepe shale further north. The sandstone is petrologically distinct from Mood-



(a)

Fig. 14 (legend p. 278)

ies sandstones on either side of the Inyoka Fault in consisting almost entirely of grains of chert and minor silicified mafic and ultramafic volcanic rocks of the Onverwacht Group and in containing less than 20% quartz (Heubeck and Lowe, 1994a). In agreement with Heinrichs (1980), we suggest that the Sibubule-Emlembe belt is made up of Fig Tree conglomerate and sandstone, not Moodies rocks (Visser, 1956; Eriksson, 1979; Jackson et al., 1987).



b)

Fault-bounded blocks of coarse- and medium-grained Moodies Group sandstone just south of the Inyoka Fault have fold axes oriented obliquely and en echelon to the Inyoka Fault, suggesting that these blocks may once have formed a single coherent block that was disrupted and imbricated by right-lateral transpression along the Inyoka Fault (Fig. 2). Sandstone composition and sedimentary structures in these blocks are similar and include planar bedding,

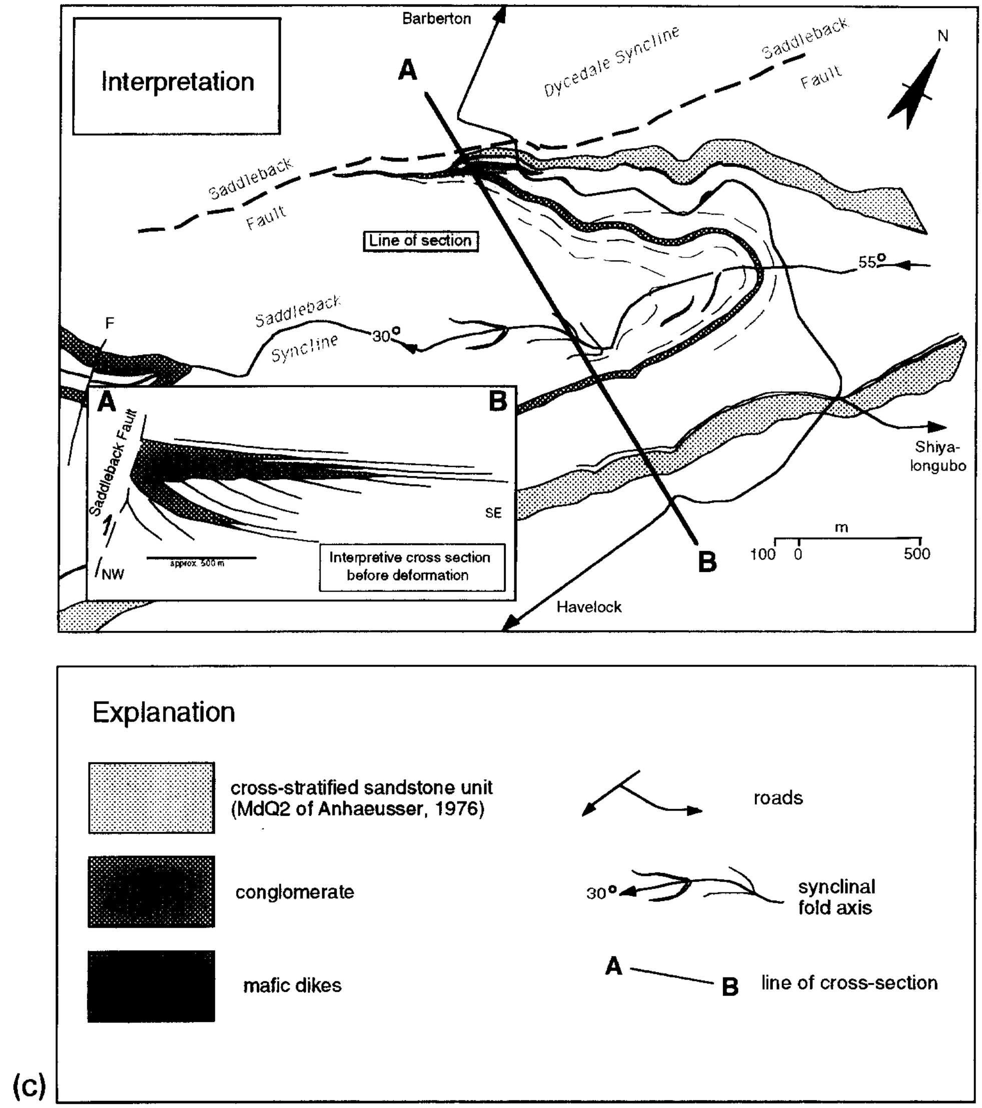


Fig. 14. Detailed features of the intraformational angular unconformity at Saddleback Pass (Fig. 13). See Fig. 2 for location. Maps illustrate southeastward tapering of the conglomerate wedge, eastward transport of sand-sized sediment interbedded with the conglomerate, and southeastward decrease of tilt of strata below the unconformity. All these attributes are consistent with the derivation of the conglomerate by up-to-the-north faulting along the Saddleback Fault.

planar and trough cross-stratification, ripples, clay drapes, and rare mudcracks, suggesting a subaerial depositional environment of variable energy. Paleocurrent data appear to be polymodal; however, this may be due to the structural complexity of this area.

The northeast-striking The Heights Syncline is strongly overturned towards the northwest and includes ~600 m of Moodies sandstone. The base of the Moodies sequence overlies and truncates strongly folded and faulted Fig Tree and Onverwacht rocks with an angular unconformity. A basal conglomerate, 0–30 m thick and dominated by clasts of plagioclase–phyric dacitic volcanic rock, is overlain by coarse-grained conglomeratic sandstone showing planar bed-

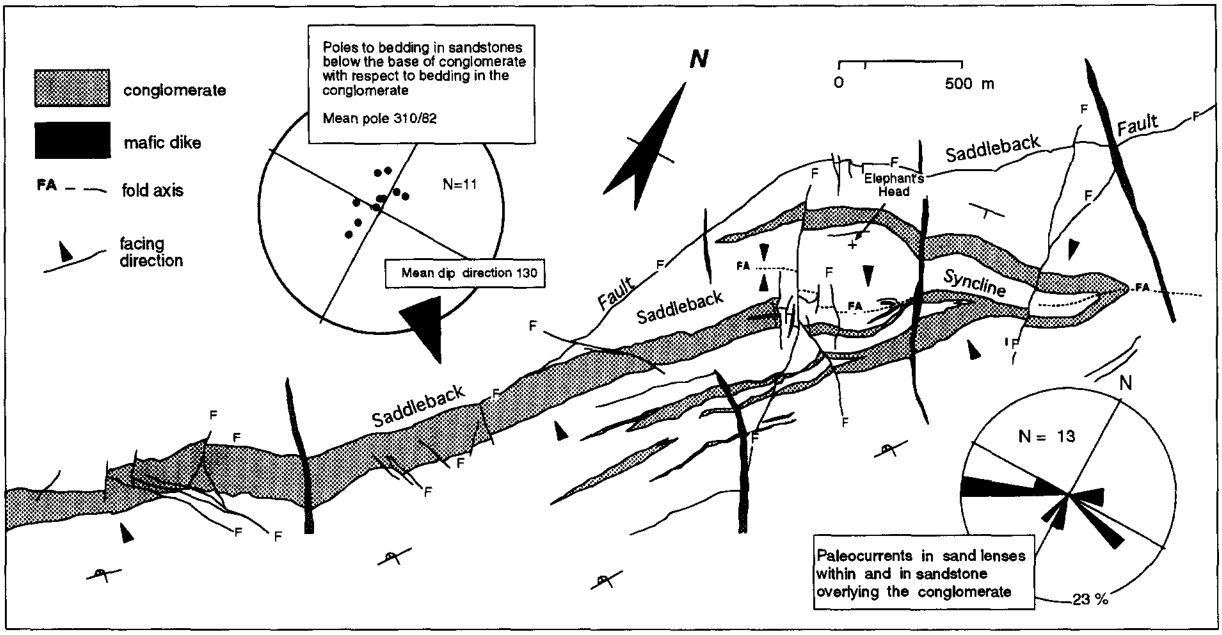


Fig. 15. Geology of the thick MdQ3 conglomerate unit at Elephant's Head, Saddleback Syncline. See Fig. 2 for location. An angular unconformity may be present at the base of the conglomerate, as suggested by the southeastward decrease in conglomerate thickness, overall southerly paleocurrents, and slightly higher dip of beds below the conglomerate (by 8° to the southeast).

ding, planar and trough cross-stratification, ripples, and rare fluid-escape structures. Paleocurrents in the central part of the belt suggest that the conglomerate was derived predominantly from the north (Fig. 16). Overlying sandstone displays a dominantly northeast–southwest bimodal paleocurrent pattern. The basal sandstone is in places cross-cut by porphyritic dikes identical in composition to Fig Tree felsic porphyries.

Shallow-water quartzose sandstone crops out in the vicinity of Piggs Peak, Swaziland, near the southeastern margin of the BGB (Fig. 1). It forms a tight, southeast-verging overturned syncline that lies within the contact-metamorphic aureole of the Mpuluzi Batholith. Deformation of conglomerate pebbles indicates both flattening and stretching strain (Lamb, 1984; Heubeck and Lowe, 1994b). The composition, inferred depositional environment, apparent unconformable relationship with underlying immature clastic rocks and ferruginous shales of the Fig Tree Group, and style of deformation suggest that these rocks are equivalents of the Moodies Group to the northwest. Rocks further south but in apparent continuity with those near Piggs Peak were assigned to the Malolotsha Group by Lamb and Paris (1988), who also described intraformational angular unconformities from this part of the belt (Fig. 1).

The stratigraphy and structure of the Moodies Group along strike towards the northeast from the central BGB are incompletely known. The tight, overturned, northwest-verging folds of the central belt south of the Inyoka Fault appear to be replaced by open, upright folds south of the Shiya Lo Ngubo Dam (Fig. 1). Moodies stratigraphy and sandstone composition resemble that of the The Heights Syncline (Heubeck and Lowe, 1994a). Further to the northeast, Moodies sandstone is in intrusive contact with and metamorphosed by the Salisbury Kop Pluton. Folded ridges of Moodies sandstone are truncated by the apparently undeformed granodioritic pluton (Heubeck et al., 1993).

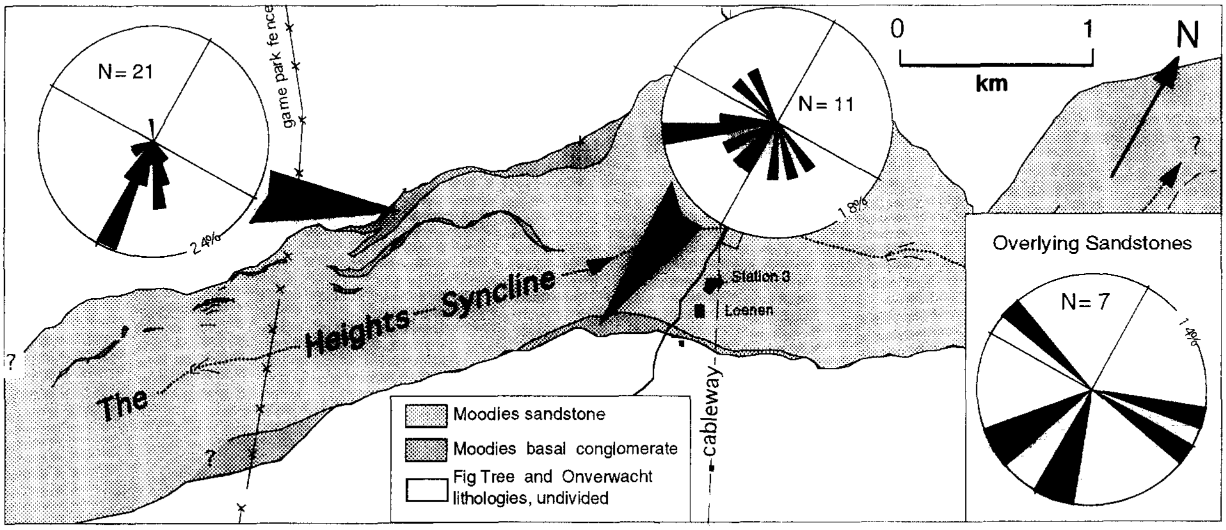


Fig. 16. Geologic map and paleocurrents of Moodies rocks in the The Heights Syncline. Southerly paleocurrents predominate in the basal conglomerate along the southern margin of the syncline. See Fig. 2 for location.

5. The Lomati River Pluton

A mafic pluton less than 1 km wide crops out over a length of ~23 km along the Inyoka Fault in the central BGB (Visser, 1956; Fig. 17). This pluton, here named the Lomati River Pluton, weathers readily and is confined in this study area largely to the valley of the Lomati River and its tributaries.

The Lomati River Pluton is composed of coarse-grained mafic plutonic rock spanning a wide compositional range. The most mafic variety comprises cumulate clinopyroxene, hornblende, calcic plagioclase, and rare iddingsite. Over most of its outcrop area, the rock consists mainly of hornblende and zoned calcic plagioclase. A leucocratic phase is composed of abundant quartz–albite graphic intergrowths and minor mafic minerals. Quartz appears to occur preferentially near the pluton margins. Most mafic minerals have been partially altered to chlorite, and plagioclase is commonly sericitized. Common zircon, sphene, and Fe-oxides, and rare sulfides form accessory minerals.

The pluton is in intrusive contact with contact-metamorphosed, baked, and recrystallized ferruginous and tuffaceous shale and greywacke of the Fig Tree Group. It truncates bedding in Fig Tree rocks at its southwestern end near the Barberton–Havelock Road on Farm Heemstede (Figs. 2, 17). Its relationship to the Moodies Group is uncertain because a contact with rocks of the Moodies Group is nowhere exposed. In Moodies outcrops close to the pluton, the basal conglomerate appears to be affected by contact metamorphism. The outcrop pattern of the Lomati River Pluton in map view suggests that it has been involved in post-Moodies deformation but the rock appears free of macroscopically or microscopically visible strain. The pluton is, however, clearly offset by late faults striking at high angles to the regional strike and also appears to be truncated by the Inyoka Fault. Few, if any, outcrops of the pluton occur south of this fault.

The northern slopes of Masenjane and the northern limb of the Maid-of-the-Mists Syncline, both localities south of the Inyoka Fault, expose a stockwork of fine-grained mafic hypabyssal rock intruding quartzose Moodies sandstone (Fig. 17). The hypabyssal rock has been completely altered to chlorite, Fe-oxides, and accessory quartz and appears massive in hand sample. Although its original fabric has been mostly obliterated, a few samples show remnant graphic texture in thin section (G. Byerly, pers.

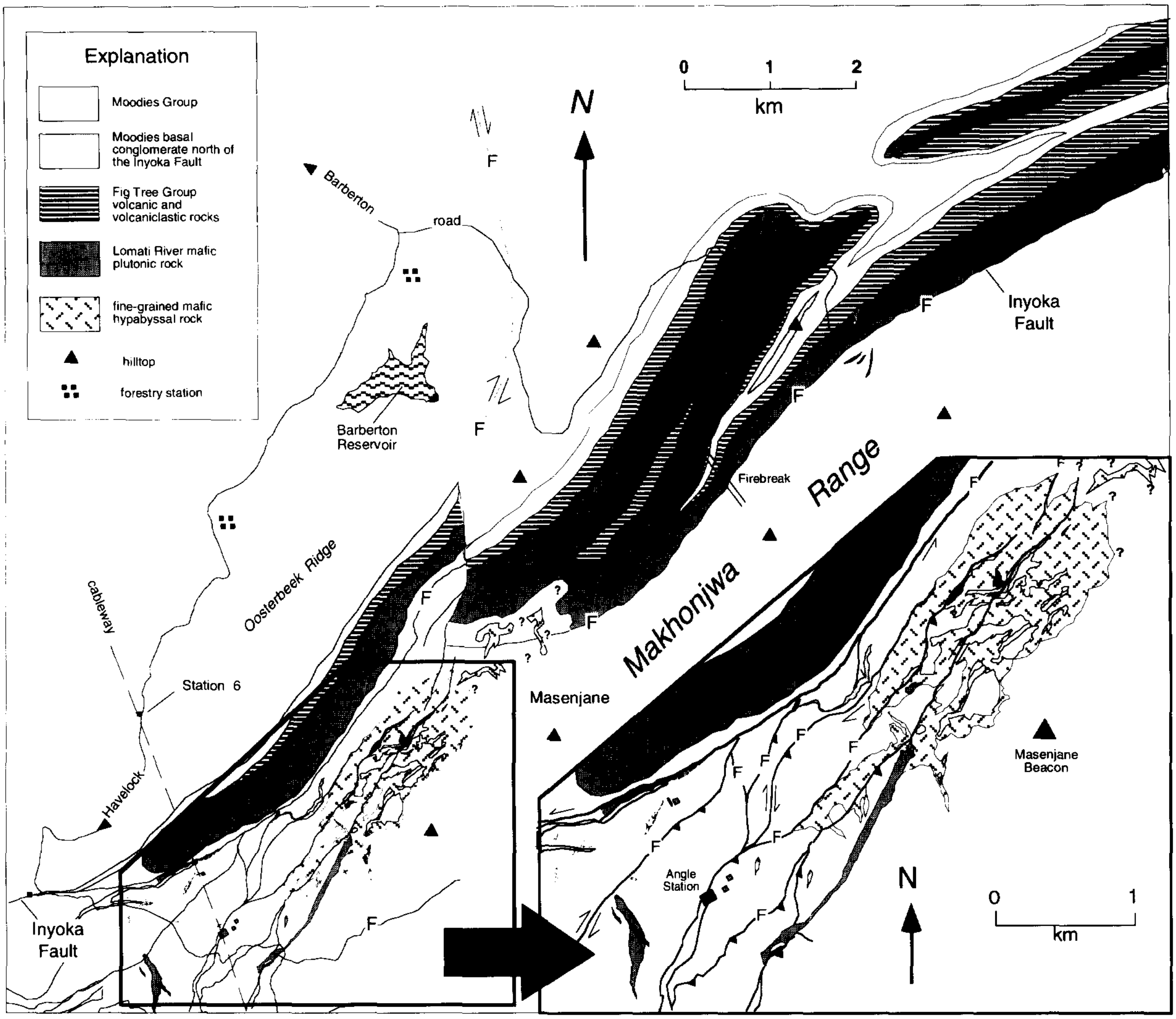


Fig. 17. Geologic map of the Lomati River Pluton along the Inyoka Fault in the central BGB. Inset map illustrates the complex relationship between marginal plutonic rocks and the stockwork of fine-grained mafic hypabyssal rocks cross-cutting Moodies sandstones south of the Inyoka Fault. See Fig. 2 for location.

commun., 1993). Detailed mapping suggests that the contact between the Lomati River Pluton and this hypabyssal metabasaltic rock is gradational and that the latter represents magma that ascended along planes of weakness generated by complex faulting at the intersection of the Inyoka Fault and several major faults striking into it at acute angles (Fig. 17).

6. Discussion

6.1. Depositional environments

The basal Moodies conglomerate was deposited as amalgamated coalescing longitudinal bars in a braided fluvial–alluvial environment (Eriksson, 1978, 1980). The overlying cross-stratified sandstone of MdQ1 was deposited on alluvial plains and channel bars, whereas the finer-grained sandstone–shale packages formed as fluvial channel-fill and overbank deposits (Eriksson, 1978). Sedimentary structures, vertical and lateral successions, and paleocurrent patterns are in good agreement with the subenvironments and depositional architecture of modern (Miall, 1978) and ancient (Miall, 1978; Vos and Tankard, 1981; Fedo and Cooper, 1991) alluvial and fluvial systems. Conglomerate from the southern BGB, thought to represent the most proximal deposits of this facies (Eriksson, 1978), is part of the Fig Tree Group and cannot be used to interpret Moodies paleogeography.

Stratigraphically higher, tidal systems, including tidal deltas, channels, flats, and sand shoals, are present in the Stolzburg, Saddleback, and Eureka Synclines (Eriksson, 1977). The overlying sequence in the Eureka and Stolzburg Synclines (upper MdS1 and MdQ2 of Anhaeusser, 1976; MD3 of Eriksson, 1979), consisting of several hundred meters of interbedded shale and banded-iron formation, coarsening upward through interbedded siltstone and sandstone to plane-stratified and rippled sandstone and capped by large-scale cross-stratified sandstone, was interpreted by Eriksson (1979) as a progradational beach sequence, including nearshore shelf, shoreface, tidal inlet, and barrier spit facies. We suggest that the upper part of this sequence, MdQ2, represents an offshore subaqueous dune field formed in an area of strong shoreline-parallel currents. This interpretation is based on the following observations.

(1) The large cross-sets are developed over more than 60 km along strike and lack evidence of changing depositional energy. Large aqueous dunes appear to have formed over a broad region, controlled by continuous, strong to moderate currents. However, a similar geometry could have also been generated by extensively migrating tidal inlet facies (Eriksson, 1979), although those facies are typically less extensive across strike (Hoyt and Henry, 1967; Galloway and Hobday, 1983).

(2) Transgressive sand sheets resulting from the reworking of tidal and deltaic deposits in shoreline environments (Eriksson, 1980) are typically thinner (1–10 m) and show planar stratification or low-angle foresets characteristic of washover deposits rather than the observed stacked large high-angle foresets (Galloway and Hobday, 1983).

(3) Large-scale cross-bedded units similar in scale, association, lateral extent, and regional paleocurrent pattern are known from migrating subaqueous dune fields worked by strong currents on shallow shelves in both recent (Harvey, 1966; Flemming, 1978) and ancient (Homewood and Allen, 1981; Blakey, 1984; Allen et al., 1985; Smith and Tavener-Smith, 1988) shelfal systems.

(4) Interpretation of MdQ2 as a migrating shelf dune field is also consistent with the presence of chemically precipitated banded-iron formation below this unit and immediately above the metabasalt capping MdQ2. The dominantly southwesterly transport direction is perpendicular to the direction of major coarse clastic input by fan deltas and therefore probably subparallel to the strike of the original depositional basin (Figs. 7 and 13). Sections of MdQ2 showing opposing paleocurrents, such as in the central Saddleback Syncline, may reflect tidal influence or counter-currents (Allen et al., 1985).

Sandstone overlying the metabasalt resembles the fluvial and marginal marine facies of MdQ1 in lithologic association and sedimentary structures. Interbedded conglomerate wedges in MdS2 and MdQ3 in the Saddleback Syncline (Figs. 14 and 15) probably formed as small fan deltas, and interbedded lenses of cross-stratified sandstone were deposited in channels during low-water flow. The two conglomerate bodies in the upper Saddleback Syncline represent discrete pulses of coarse-sediment progradation over shallow shelf (Eriksson, 1978) or tidal deposits. The fining-upward sequence overlying the upper conglomerate (MdQ3) in the Saddleback Syncline records the rapid re-establishment of deeper-water shelfal conditions.

6.2. Syndepositional deformation

The geometry of the conglomerate wedges in the upper Moodies Group, the orientation of upturned bedding unconformably below the conglomerates, paleocurrent directions, and the sense of displacement along the Saddleback Fault all suggest that the fan deltas in the upper Moodies Group of the Saddleback Syncline developed as a response to up-to-the-north faulting along one or several of the major faults within and/or bordering the BGB to the north (Figs. 14 and 15). The intraformational angular unconformity is best interpreted as having formed as fault-dragged bedding planes were truncated by a prograding fan delta. The progressive grading of the unconformity laterally into a conformity and the lack of evidence for a depositional hiatus suggest that faulting and its sedimentary response occurred contemporaneously. The setting of the unconformity is similar to well-documented examples from the southern Pyrenean foreland basin (Anadón et al., 1986).

The sense of fault displacement and the direction of shortening correspond to orientation and vergence of the regional late Moodies to early post-Moodies synclinal fold-and-thrust belt in the central BGB (Figs. 1 and 2; Heubeck and Lowe, 1994b). The stratigraphic position of the unconformity and of the fan-delta wedges near the top of the Moodies Group suggests that they represent early responses to deformation that continued in post-Moodies time. The shortening of the Moodies Group by regional folding, combined with the evidence for contemporaneous up-to-the-north faulting, suggests that this deformational event involved northwest-to-southeast-directed shortening. We therefore suggest that the upper Moodies Group north of the Inyoka Fault was deposited in front of southeast-verging thrust faults that were forming along the present-day northern margin of the BGB.

Stratigraphic correlation of the Malolotsha Group, a likely Moodies Group equivalent in Swaziland, suggested to Lamb and Paris (1988) that these rocks, as well as the deformation event that deformed them, became younger towards the southeast. This coincides with the propagation direction of the fold-and-thrust belt of late Moodies time described here. However, fold vergence and sense of fault displacement between the southern margin of the BGB and the Inyoka Fault also indicate that shortening was directed towards the northwest, that is, towards the interior of the BGB (Lamb, 1984; Paris, 1985; de Ronde and de Wit, 1994; Heubeck and Lowe, 1994b). Because the stratigraphic correlation between Moodies rocks north of the Inyoka Fault and the Malolotsha Group is still tenuous, a causal relationship between these two similar styles of deformation acting on rocks of possibly identical age does not yet appear to be possible.

6.3. System architecture

Depositional environments of the Moodies Group north of the Inyoka Fault record an overall transgressive sequence punctuated by the depositional effects of syndepositional faulting late in the basin history (Fig. 18). Moodies deposition began with a basal alluvial conglomerate that grades upwards into braided fluvial facies of MdQ1 (Fig. 6). This is succeeded by a fining-upward succession of deltaic, tidal, and shoreline environments into moderately deep marine facies in MdS1 (Fig. 18). Local abrupt shallowing associated with syndepositional faulting records the onset of basin shortening in the Saddleback Syncline whereas environments further north, in the Moodies Hills and the Eureka Syncline, apparently record more-or-less continuous subaqueous environments (Fig. 18).

Correlation of higher parts of the Moodies Group among the various blocks north of the Inyoka Fault remains problematic. Rapidly shifting transport patterns suggest that the depositional systems responded quickly by reorientation to only subtle changes in depositional parameters, such as shoreline trend and position, direction and magnitude of sediment supply, fluctuations in longshore currents, location of sediment input, and water depth (Figs. 9a and 9b). It also shows that paleocurrents in the Moodies Group must be interpreted on a bed-by-bed basis.

Depositional environments south of the Inyoka Fault appear to be limited to terrestrial settings. Alluvial and braided stream facies dominate; no marine units were identified. The lack of distinctive lithologic units prevents the correlation of these rocks with those north of the In-

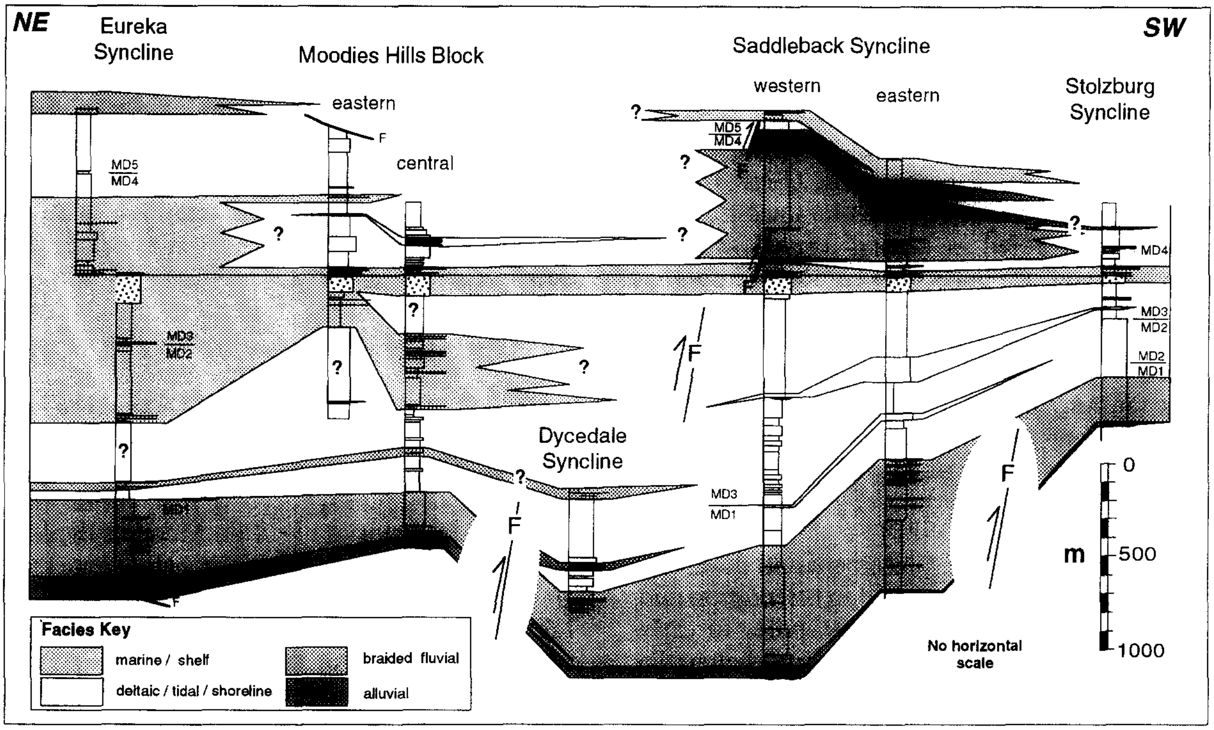


Fig. 18. Proposed facies correlation diagram of the Moodies Group north of the Inyoka Fault, compiled from lithologic and facies descriptions by Reimer (1967), Eriksson (1977, 1978, 1979, 1980), Hose (1990), and this study. Moodies Group subdivisions of Eriksson (1980) are shown and can be correlated to Anhaeusser's (1976) subdivisions using Fig. 5. Patterns of stratigraphic units are defined in Fig. 4. Datum is the widespread second basaltic marker horizon (MdL2). The lower Moodies Group consists of a fining- and deepening-upward sequence ranging from a basal alluvial to a capping shelf facies. The upper Moodies Group is punctuated by at least two pulses of shallowing high in the section of the Saddleback and possibly Stolzburg Syncline, related to intraformational angular unconformities and the onset of basin shortening.

yoka Fault, although their lithologies and facies sequence generally resemble those of the lower Moodies Group. The lack of K-feldspar and only rare occurrence of altered plagioclase, however, suggests that the rocks south of the Inyoka Fault had a different provenance from those to the north and may have been deposited in separate basins.

6.4. *Provenance of Moodies sands*

Moodies paleocurrent patterns from individual units can be grouped into two classes. Conglomeratic units in the upper Moodies Group consistently show a southerly transport direction (Fig. 7). These conglomerates are interpreted to represent south-facing fan deltas shed from active fault scarps by up-to-the-north faulting along one or more of the major strike-parallel faults along the northern margin of the central BGB (Figs. 7, 14, 15). Paleocurrents in sandstone not associated with conglomerate, in contrast, indicate northerly, southerly, or bidirectional transport directions, mostly subparallel to the strike of the orogen and perpendicular to the direction of shortening (Fig. 7). We interpret these patterns, most of which are recorded from marginal marine and shelf environments, to reflect reworking of originally northerly derived sediment by shoreline-parallel tidal and shelfal currents.

Paleocurrent patterns in conglomeratic units, facing direction of the fan-delta wedges in the Saddleback Syncline, decrease in feldspar as sandstone framework component in Moodies sandstones from NE to SW, and up-to-the-north sense of displacement along most of the major faults along the northern margin of the central BGB all suggest that a significant amount of Moodies detritus was delivered from the north (Heubeck and Lowe, 1994a). This interpretation is consistent with the structural style of the central BGB (Daneel, 1987; Heubeck and Lowe, 1994b) north of the Inyoka Fault but contrasts with the inferred provenance of the Moodies Group developed by Eriksson (1979) and the tectonic models of Fripp et al. (1980) Jackson et al. (1987) and de Wit et al. (1992).

6.5. Tectonic setting

A number of features of the lower Moodies Group on both sides of the Inyoka Fault suggest that deposition may have occurred in an extensional setting. These include:

(1) The presence of a thin, widespread basal alluvial conglomerate transgressive over largely undeformed volcanic and volcaniclastic strata of the upper Fig Tree Group and over deformed lower Fig Tree and Onverwacht Group strata is consistent with early sedimentation patterns in extensional basins (Leeder et al., 1988; Busby-Spera, 1988; Martini and Sagri, 1993). The local provenance of clasts in the conglomerate suggests that the source was not a distant fold-and-thrust system.

(2) Thick fining- and deepening-upward fluvial to marginal marine facies trends in MdQ1 and MdS1 suggest that subsidence and source-area uplift continued without substantial progradation of an orogenic front.

(3) Interbedded volcanic and volcaniclastic strata, including the metabasaltic units MdL1 and possibly MdL2, and several felsic tuff horizons are consistent with volcanic styles in extensional settings (Neumann and Ramberg, 1978; Burke and Kidd, 1980; Miall, 1984).

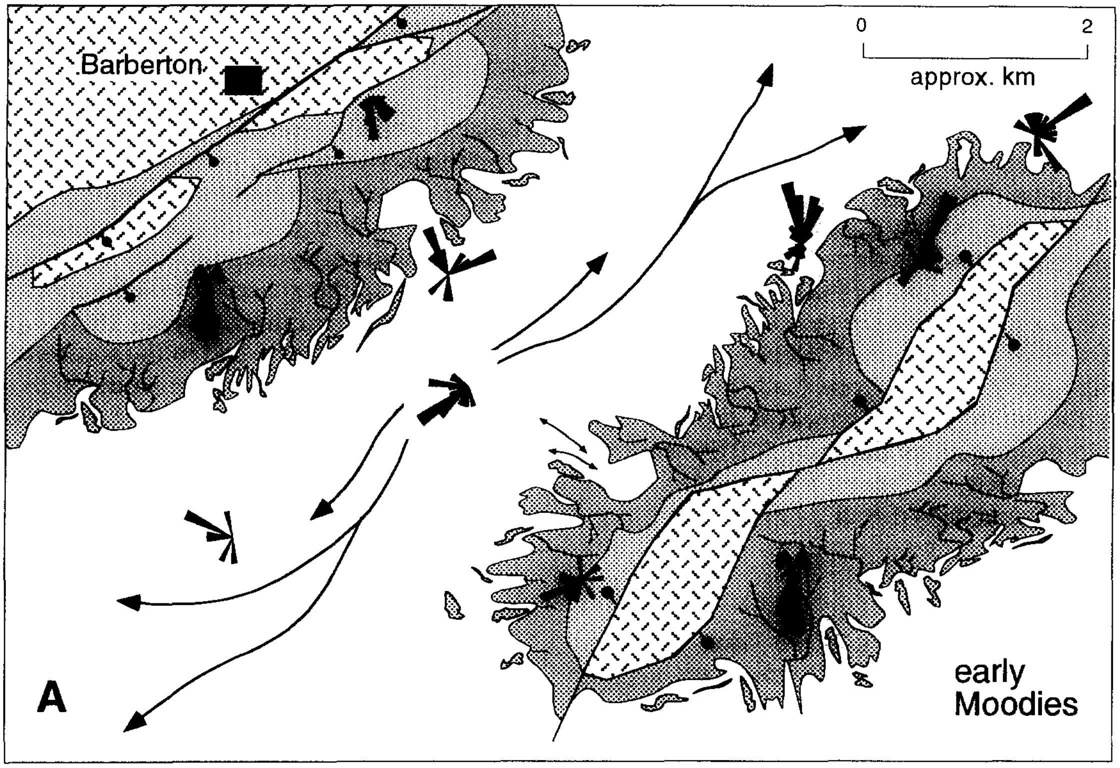
These features are similar to those of post-orogenic extensional or strike-slip basins developed on Phanerozoic orogens (Arthaud and Matte, 1977; Biddle and Christie-Blick, 1985; Etheridge, 1986; Ziegler, 1990).

The upper Moodies Group, above the metabasaltic marker, shows a number of attributes consistent with deposition in a foredeep. The intraformational angular unconformity in the upper Moodies Group of the Saddleback Syncline reflects incipient syndepositional shortening by a south- or southeastward advancing thrust front along the northern margin of the BGB. Continued shortening is recorded in the southeast-vergent fold-and-thrust belt of the central BGB north of the Inyoka Fault (Heubeck and Lowe, 1994b). Paleocurrent patterns throughout the Moodies Group suggest extensive reworking of clastic material by regional currents flowing parallel to the basin axis; these recycling-intensive marginal marine environments are most widespread in the upper Moodies Group. The common quartzose sandstone clasts and the variety of intra-greenstone belt lithologies in the south-facing fan-delta wedges also attest to the increased role of sedimentary recycling and to uplift of greenstone belt rocks (Heubeck and Lowe, 1994a).

The tectonic setting of the Lomati River Pluton and the associated mafic hypabyssal rocks cannot yet be resolved from the available geologic and petrographic data. The intrusion of the pluton may be related to the earlier extensional phase of the Moodies basin, and its hypabyssal mafic equivalents could possibly be feeders to the Moodies metabasalt MdL2. Alternatively, the pluton could also be a result of basal crustal melting following tectonic shortening, thickening, and loading of the crust. Further analytical and geochronological data are necessary to resolve the geologic history of the Lomati River Pluton.

6.6. Paleogeography

Paleogeographic reconstructions of the Moodies basin are difficult because of the steep dip of bedding throughout the central BGB, the paucity of unambiguous time lines, and the poorly constrained amount of shortening and strike-slip displacement on major faults. Facies changes normal to the strike of the depositional basin are only poorly constrained. Due to these difficulties, the paleogeographic maps shown in Fig. 19 must be considered speculative. While we agree



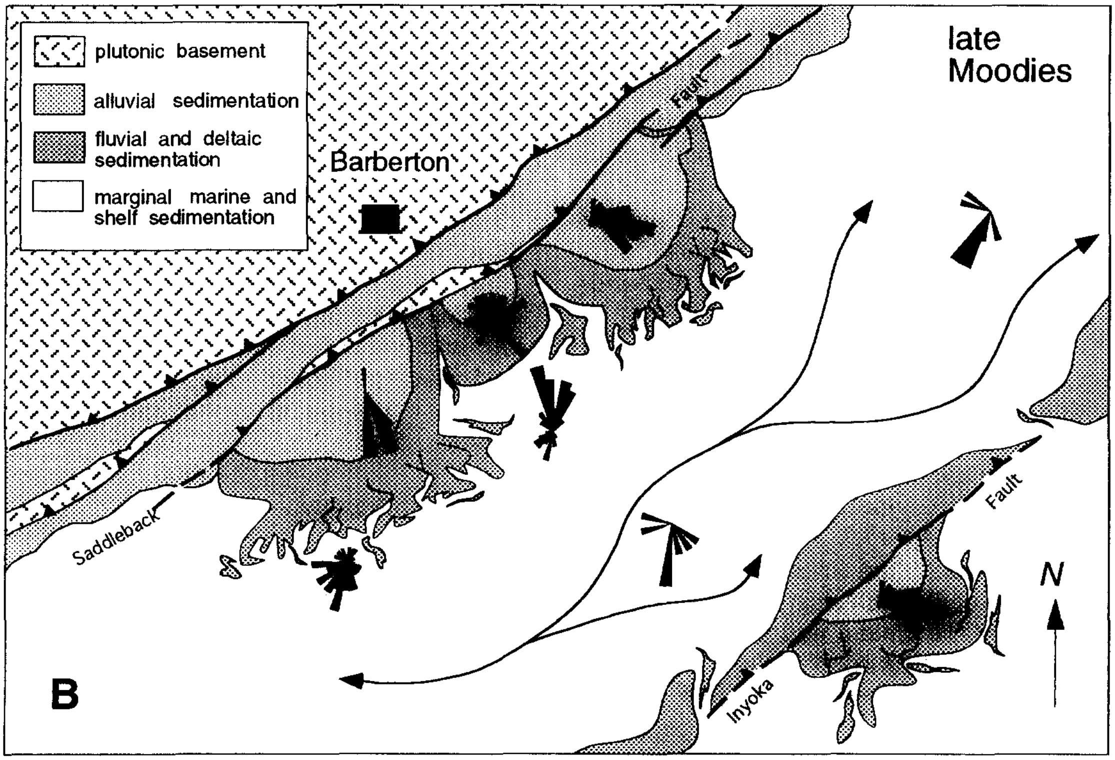


Fig. 19. Schematic paleogeographic reconstructions of the early and late Moodies basin combined with paleocurrent rose diagrams. (A) The lower part of the Moodies Group, MdB through MdQ1, was deposited as alluvial and fluvial deposits flanking uplifts bounded by normal faults. These terrestrial settings were succeeded by marginal marine and open marine systems. Principal clast input was from the north but intra-basinal uplifts may have been important. Sediment transport was dominated by shoreline-parallel currents. (B) Basin inversion of late Moodies time by southeast-directed thrust faulting along the northern margin of the BGB led to basin-margin uplift, recycling of greenstone belt material, and the development of local angular intraformational unconformities.

with most of the environmental interpretations of Eriksson (1978, 1979, 1980), our reconstruction differs in including: (1) a predominant northern provenance of Moodies Group rocks; (2) a two-stage tectonic evolution involving early extensional and late compressional Moodies basins; and (3) syndepositional deformation and shortening along an orogenic front to the northwest in late Moodies time.

7. Conclusions

Lower Moodies rocks north of the Inyoka Fault form a fining- and deepening-upward clastic sequence representing an upward transition from alluvial to shallow shelf environments. Basin deepening was probably related to subsidence caused by extension of an orogenic basement that formed in late Fig Tree time. Sedimentology and sandstone petrography of the Moodies Group indicate extensive reworking in tidal and shallow-marine environments of mostly sand-sized material that originated largely in felsic plutonic rocks located north of the greenstone belt. Rare mafic volcanic rocks and felsic tuffs are interbedded with the clastic sequence and suggest at least local extrusive igneous activity during early basin-fill. Syndepositional thrust faulting on southeast-verging faults along the northern margin of the belt in late Moodies time marked initial basin collapse and shortening. Detritus derived by erosion of the uplifted greenstone belt was recycled in fan deltas prograding south from fault scarps. A foredeep setting is consistent with sedimentation patterns and the style of subsequent deformation.

Moodies rocks south of the Inyoka Fault appear to have been deposited in terrestrial environments. They may be time-correlative with the lower Moodies Group north of the Inyoka Fault but lack preserved feldspar. The mafic Lomati River Pluton and its hypabyssal equivalents appear to have intruded deformed Moodies rocks on both sides of the Inyoka Fault.

The continuity and apparent rapidity of Moodies sedimentation suggest that deposition may have been completed in only a few million years. This hypothesis is supported by indirect geochronologic constraints.

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