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Evolution and Subsidence of Early Precambrian Sedimentary Basins [and Discussion]

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Evolution and subsidence of early Precambrian sedimentary basins

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Many of the models for modern sedimentary basins postulate two-stage subsidence; a rapid initial subsidence due to thinning or loading of the crust, followed by a more protracted thermal stage as the lithosphere, which is thinned during the initial stage, relaxes to equilibrium thickness. The geology of a number of Archaean greenstone belts and early Proterozoic cratonic basins in South Africa may be explained by such a model. Rapidly erupted shallow marine or subaerial volcanic rocks predominate in the lower parts of sedimentary–volcanic sequences. These are thought to relate to initial subsidence as (1) accommodation of relatively thick volcanic sequences requires substantial and rapid subsidence, and (2) marginal uplift following these early volcanic intervals is consistent with viscous relaxation following the initial elastic response of the lithosphere to localized loading. Sedimentary sequences overlying initial volcanic dominated intervals may have been deposited during the ensuing phase of more widespread subsidence related to thermal relaxation of the thinned lithosphere. If so, sediment-filled subsidence of *ca.* 5.5 km in greenstone terrains at 3.5 and 2.6 Ga and of 7–10 km in cratonic shelf basins between 2.7 and 2.1 Ga require increases of lithosphere thickness between *ca.* 60 and 90 km. These minimum estimates of early Precambrian lithosphere thickness, although crude, are similar to estimates of present lithosphere thickness. In some early Precambrian basins, the cause of subsidence may have been crustal extension with development of faulted grabens that evolved into continental margins, but in cratonic shelf basins faulting did not occur during or after the initial subsidence, and some less obvious causal mechanism must be sought.

Introduction

Mechanisms for sedimentary basin formation and preservation are still both controversial and puzzling. A variety of mechanisms have been proposed (see, for example, Sleep *et al.* 1980), but the similarity of rates of subsidence to those expected from thermal relaxation of a thinned lithosphere suggests that the main tectonic control is thermal and on the scale of the lithosphere (see, for example, Sleep (1971), Watts & Ryan (1976), Steckler & Watts (1978), Keen (1979), Haxby *et al.* (1976), Sclater & Christie (1980) and Sleep *et al.* (1980)). This mechanism is the same as the subsidence of the ocean floor where, with increasing age, the lithosphere cools, becomes denser and sinks to maintain isostatic equilibrium.

McKenzie *et al.* (1980) suggested that the amounts of thermally controlled subsidence in Precambrian sedimentary basins might allow estimates of lithospheric plate thickness, estimates not easily derived from other evidence but of critical significance to interpretation of Archaean tectonic processes (Bickle 1978; Burke & Kidd 1978). In this paper, we re-examine models for sedimentary basin formation dependent on subsidence of a cooling lithosphere to identify those aspects relevant to the early Precambrian. This is followed by a discussion of a few of the better studied early Precambrian basins, to analyse the genetic significance of different facies associations in terms of the different models for basin formation, and to evaluate the significance of sedimentary thicknesses in terms of lithosphere thickness.

Sedimentary basin formation: models

A permanent, isostatically compensated, sedimentary basin will only be preserved on continental crust if the crust is thinned, or if the density of the crust or underlying lithosphere is changed. Permanent density changes without associated compositional changes within the lithosphere require special pleading, especially as critical garnet-producing reactions are probably near isobaric (O'Hara *et al.* 1971; Newton 1977).

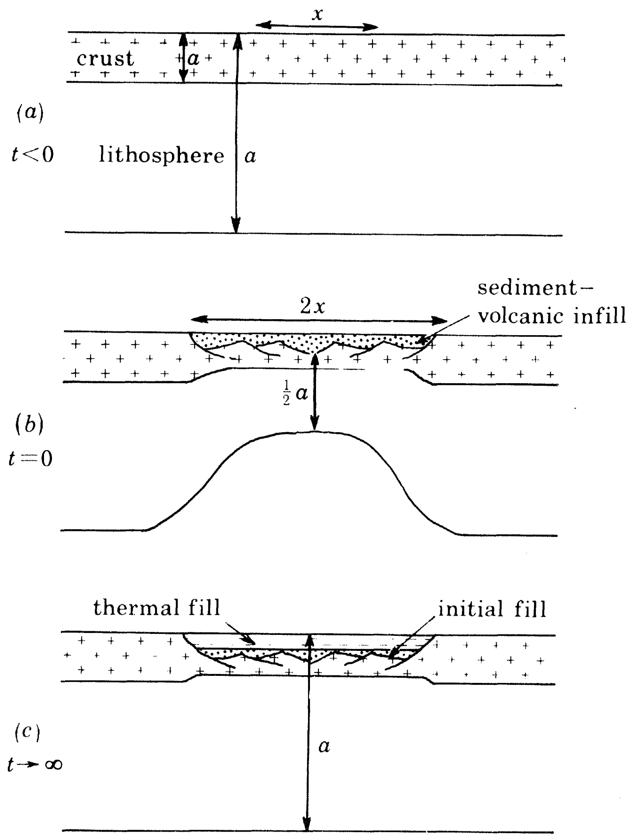

Figure 1. Schematic evolution of a sedimentary basin following crustal extension, after McKenzie (1978). (*a*) Before stretching. (*b*) Initial subsidence. Sediment or volcanic fill, or both, deposited during faulting of basement. (*c*) Thermal subsidence related to relaxation of lithosphere to original thickness. Any mechanism that rapidly thins the lithosphere and thins or loads the crust will give a similar pattern of subsidence.

McKenzie (1978) has examined the tectonic implications of stretching continental crust and underlying lithosphere, and the evolution of several modern sedimentary basins can be successfully explained by this model (see, for example, Sclater & Christie 1980) even though available data do not necessarily exclude alternative models. McKenzie's model is essentially simple (figure 1). Continental crust and lithosphere in thermal equilibrium are stretched during a rifting phase (the lithosphere is defined as a thermal boundary layer). Subsidence will occur in two stages. The first, the 'initial stage', which may also result in uplift, results from the immediate isostatic compensation of thinned crust and lithosphere, and the second, the 'thermal stage', involves subsidence at an exponentially decreasing rate as the lithosphere cools and returns to equilibrium thickness.

McKenzie *et al.* (1980) show how initial subsidence and total subsidence depend on the crustal thickness, amount of stretching, and lithosphere thickness. Total subsidence depends only on the amount of crustal thinning and the density of the sedimentary fill. The thermal subsidence depends only on the increase in lithosphere thickness after the initial stretching.

Alternative models for basin development that postulate relatively rapid crustal thinning, or effective crustal thinning with associated thinning of the lithosphere, can result in rather similar subsidence histories. For example, rapid dyke injection and lithosphere thinning, a phase change triggered by a hot mantle diapir, or even rapid erosion with an associated thermal event, will all result in rapid isostatic adjustment to the new crustal thickness or density followed by the slower thermal adjustment of the lithosphere. In these cases, a direct relation between crustal and lithospheric thinning will not be be attained, as it is in the rifting model. Even with rifting, if there is associated volcanicity, the heat transported from depth might thin the lithosphere destroying the simple relation.

For a large class of basin formation models, the subsidence history can therefore be divided into an initial phase directly relating to the mechanism causing crustal subsidence and a thermal phase relating to thermal relaxation of the thinned lithosphere. The discussion of early Precambrian basins below will be concerned with two main aspects: (1) separation of the initial and thermal subsidence, and (2) estimates of subsidence during both the initial and the thermal phases. Separation of initial and thermal subsidence can only be attempted on the basis of the interpretation of sedimentary–tectonic environments, and these in turn should place some constraints on the fundamental cause of basin formation.

Early Precambrian sedimentary basins

Early Precambrian tectonic processes, and consequently the setting of the sedimentary sequences, are much debated with little or no general consensus. We propose to discuss the evolution of early Precambrian sedimentary basins in terms of three main sedimentary–tectonic associations recognized in well preserved and well studied areas. Additional associations undoubtedly exist, and the affinities of many well known areas are not yet clear. This simple subdivision provides a convenient way to present a discussion of basin evolution in terms of phenomena that are at least adequately documented and were apparently relatively wide-spread: if basin development can be related to some simple physical model, the recognition of re-occurring features must initiate elucidation of the model.

The three associations are: (1) Archaean greenstone associated sediments, which probably encompasses more than one environment; (2) cratonic shelf basins, such as the Witwatersrand and Transvaal supergroups in South Africa; and (3) cratonic rift basins, of which an example is the Ventersdorp in South Africa.

Archaean greenstone association

All of the well preserved sedimentary sequences older than 3000 Ma, and many of the sequences formed between 2500 and 3000 Ma, occur in the Archaean greenstone association.

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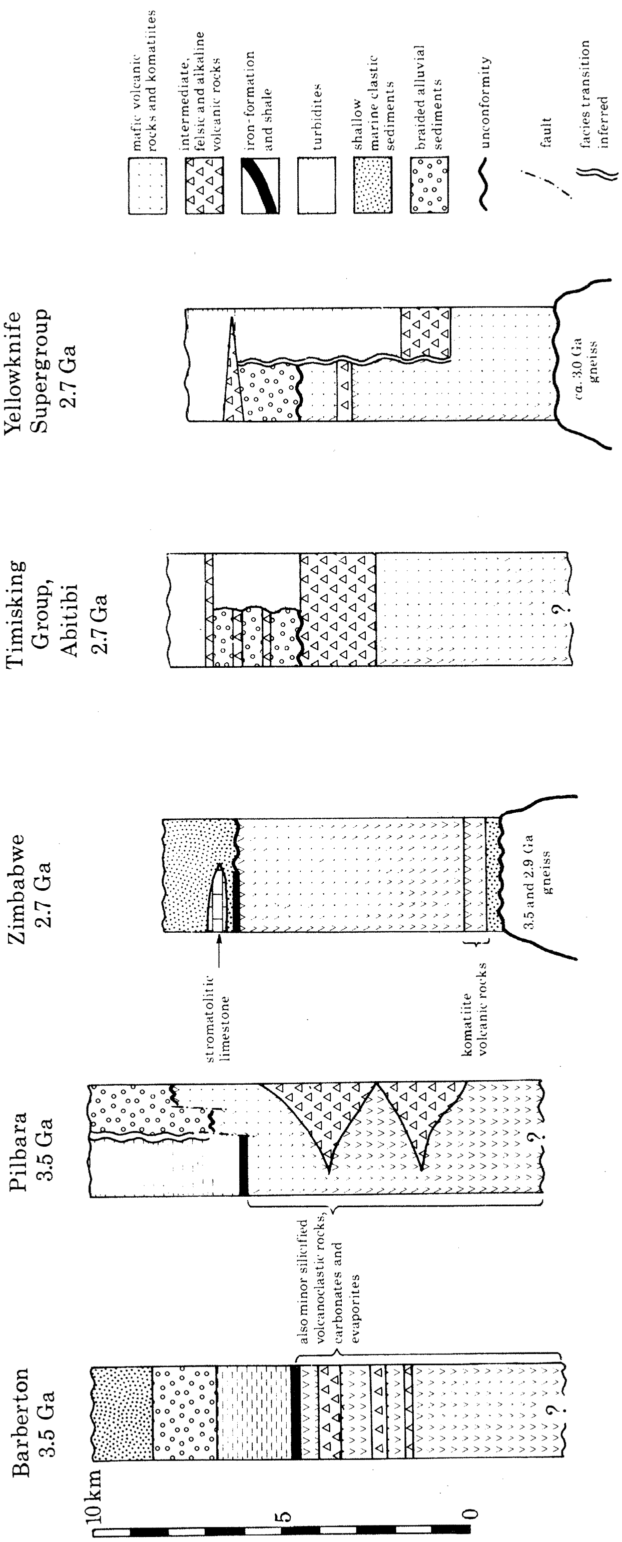

FIGURE 2. Schematic greenstone belt stratigraphic sections (based on references given in table 1).

Many greenstone belts are characterized by thick mafic volcanic sequences with subsidiary felsic and ultramafic volcanics, which are overlain by major sedimentary units. The volcanics and sediments are preserved between domal or elongate–domal granite–gneiss batholiths, which dominate the structural setting.

(*a*) *Tectonic setting of Archaean greenstone belts*

A variety of sedimentary–tectonic environments is recognised in greenstone terrains. Interpretation of virtually all greenstone areas is frustrated by uncertainty as to the nature of the basement to the volcanic sequences, especially whether the basement was a mafic–oceanic or granitic–continental type crust (Anhaeusser 1973; Hunter 1974), together with stratigraphic and structural complexities. In a few areas, geochronology and mapped unconformities have proven continental basement, but in most areas the nature of the basement is unknown, being obscured by deformation, metamorphism and granitic plutonism. Despite these problems study of sedimentary facies relations in a number of Archaean areas is sufficiently advanced to provide insight into Archaean sedimentary–tectonic processes. These studies have led to the recognition of continental margin sequences in both the *ca.* 3500 Ma Barberton Mountain Land (South Africa) and Pilbara (Western Australia) greenstone belts, continental shelf sequences in the *ca.* 2600 Ma Zimbabwe greenstone belts and volcanic arc related depositional environments in the Canadian Superior and Yellowknife provinces.

(*b*) *Basin evolution*

Among the best preserved and exposed greenstone sequences are the relatively old 3500 Ma Barberton Mountain Land and Pilbara Block. In both areas, a thick volcanic sequence is overlain by the main sedimentary sequence (figure 2, table 1). The intensity of deformation and metamorphism is laterally variable. Parts of the stratigraphy are preserved in simple synclinoria or even as flat-lying undeformed units. Elsewhere, particularly in the lower parts of the volcanic sequence, deformation and metamorphism are more intense.

The volcanic units (Onverwacht Group in Barberton, Warawoona Group in the Pilbara) range from komatiitic to felsic volcanics, although pillowed komatiitic and tholeiitic basalts predominate. Thin and often cherty metasediments are interbedded with the volcanic rock. These represent silicified carbonates, volcanoclastic sediments and evaporative sulphates (Barley *et al.* 1979; Lowe & Knauth 1977; Dunlop & Buick 1981). This sedimentary evidence indicates that the volcanic rocks accumulated in very shallow marine environments. It is impossible to give precise estimates of the thicknesses of the volcanic units in the Pilbara and at Barberton, as the bases of the sequences are not preserved. In the Pilbara, relatively undeformed sections 5–8 km thick may be measured (Barley *et al.* 1979), and gravity anomalies preclude much greater thicknesses. In Barberton, estimates of 14 km by Viljoen & Viljoen (1969) may include unrepresentative sections, and do not take into account potential structural complexities (Williams & Furnell 1979).

The volcanic sequences are overlain by thick clastic sedimentary units in both areas, and Eriksson (1981 *a*) has constructed rather similar palaeogeographic models in both Barberton and the Pilbara (figure 3). Laterally persistent iron-formation–shale units overlie areas of the volcanics, suggesting early rapid subsidence of the shallow water volcanic sequence. Thick and aerially extensive turbidites pass either laterally or vertically into braided alluvial sequences. In the Pilbara, this transition is entirely lateral and abrupt, with no intervening

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Table 1

			initial phase			thermal phase		
basin	basement	age and duration	rock types	thickness	density assumed	rocks types	thickness	density assumed
Barberton, S. Africa	not known; sialic crust inferred from sediments	3540[1] ± 30 Ma; duration not known	komatiitic, tholeiitic and felsic volcanic rocks, carbonate, minor volcanoclastics, chert, evaporites: shallow marine[15, 16]	not known	2.9	iron-formation, shale, turbidites overlain by braided alluvial and shallow marine[17]	5.5 km	2.55
Pilbara, W. Australia	not known; sialic crust inferred from sediments	3556[2] ± 32 Ma; duration not known	komatiitic, tholeiitic and felsic volcanics. Minor volcanoclastics, silicified carbonate, evaporites: shallow marine[18]	> 5 to 8 km	2.9	iron-formation, shale overlain by turbidites or braided alluvial[19]	max. *ca.* 5 km; 3 km of turbidites with water depth $\gtrsim$ 500 m	2.55
Pongola, S. Africa	3160[3] Ma granites	3090[4] ± 90 Ma on volcanics; intruded by 2819[5] Ma mafic dyke	basal arkosic, arenites, mainly tholeiitic basaltic lavas, subordinate stromatolitic dolomites and shale: shallow marine[20]	5.5 km	2.7	shale, siltstones, conglomerates: intertidal + shallow marine[20, 21]	3 km	2.55
Younger greenstones, Zimbabwe	3500[6] and 2900[7] Ma granite, gneiss and greenstones	2640[8] ± 140 Ma; intruded by 2570[33] ± 15 Ma granites	basal shallow marine and intertidal sediments; mainly komatiites and tholeiitic basalts: shallow marine[22, 23]	5–7 km	2.9	conglomerates, siltstones, minor iron-formation stromatolitic limestones, quartzite: shallow marine[22, 24]	2 km	2.55
Abitibi Supergroup & Timisking Group, Abitibi Greenstone Belt, Canada	not known	*ca.* 2700[9] Ma	mafic and intermediate lavas: mainly submarine[25]	not known	2.9	turbidites overlie and are laterally equivalent to braided alluvial; minor volcanic horizons[25]	3.5 km turbidites at top in water depths $\gtrsim$ 500 m	2.55

Table 1 (*cont.*)

basin	basement	age and duration	initial phase: rock types	initial phase: thickness	initial phase: density assumed	thermal phase: rock types	thermal phase: thickness	thermal phase: density assumed
Yellowknife Supergroup, Slave Province, Canada	*ca.* 3000 Ma granitic gneiss[10]	2665[11] ± 15 Ma on volcanic rocks; intruded by 2500 Ma granites[11]	mainly tholeiitic submarine basalts, minor felsic volcanics[10]	< 10 km	2.9	turbidites laterally equivalent to and overlie braided alluvial sediments; minor felsic volcanics[26]	5.5 km turbidites at top of section. Water depths > 500 m	2.55
Witwatersrand Supergroup, S. Africa	*ca.* 2800[34] Ma granitic rocks	*ca.* 2700 Ma; overlain by 2643[12] ± 80 Ma lavas	basal braided alluvial subaerial basaltic volcanic rocks[27]	2 km	2.83	intertidal quartz arenites and shales overlain by braided alluvial arkoses and conglomerates[27]	7 km	2.55
Ventersdorp Supergroup, S. Africa	2700[3] Ma granitic crust and Witwatersrand sequence	2643[12] ± 80 Ma; overlain by 2240 Ma Transvaal Supergroup	subaerial basaltic volcanic rocks, alluvial fan and lacustrine sediments related to graben development[28,29]	3 km	2.75	not well defined; braided alluvial overlain by basic volcanic rocks[28,29]	500 m	2.55
Transvaal Supergroup, S. Africa	2480[13] Ma granites	2200[5,35] Ma on volcanic interval; cut by Bushveldt, 2100[13,14] Ma	braided alluvial, basaltic volcanic rocks, intertidal argillite, arenite[30]	2 km	2.66	shallow marine and intertidal dolomites overlain uncomformably by shallow marine arenites and shales[31,32]	10 km	2.66

References: (1) Hamilton *et al.* (1979), (2) Hamilton *et al.* (1980), (3) Allsopp *et al.* (1962), (4) Burger & Coertze (1973), (5) Davies *et al.* (1970), (6) Moorbath *et al.* (1977), (7) Hawkesworth *et al.* (1979), (8) Hamilton *et al.* (1977), (9) Goodwin (1977), (10) Padgham (1980), (11) Henderson (1981), (12) Van Niekerk & Burger (1978), (13) Coertze *et al.* (1978), (14) Hamilton (1977), (15) Viljoen & Viljoen (1969), (16) Lowe & Knauth (1977), (17) Eriksson (1980), (18) Barley *et al.* (1979), (19) Eriksson (1981*b*), (20) Mathews (1967), (21) von Brunn & Hobday (1976), (22) Bickle *et al.* (1975), (23) Nisbet *et al.* (1977), (24) Martin *et al.* (1980), (25) Hyde (1980), (26) Henderson (1972, 1975), (27) Pretorius (1976), (28) Winter (1976), (29) Buck (1980), (30) Button (1973*a*), (31) Eriksson *et al.* (1976), (32) Button & Vos (1977), (33) Hickman (1978), (34) Allsopp (1964), (35) Button (1976).

shallow marine shelf facies yet recognized. In Barberton, turbidites of the Fig Tree Group pass vertically into fluvial sediments of the Moodies Group. Eriksson (1981*a*) interprets this relation in terms of a steep rift-type continental margin, with alluvial sediments prograding directly onto fan-slope deposits (figure 3). Shallow marine facies only become well developed in the Moodies Group higher in the stratigraphy at Barberton, after the establishment of a continental shelf. The scale of the turbidite deposits in the Pilbara (tens of thousands of square kilometres) and evidence for a substantial tidal range at Barberton have been taken to suggest that these basins were marginal to major oceans. The sediments in both areas contain a high proportion of potash feldspar. Eriksson (1981*a*) cites this as evidence that a

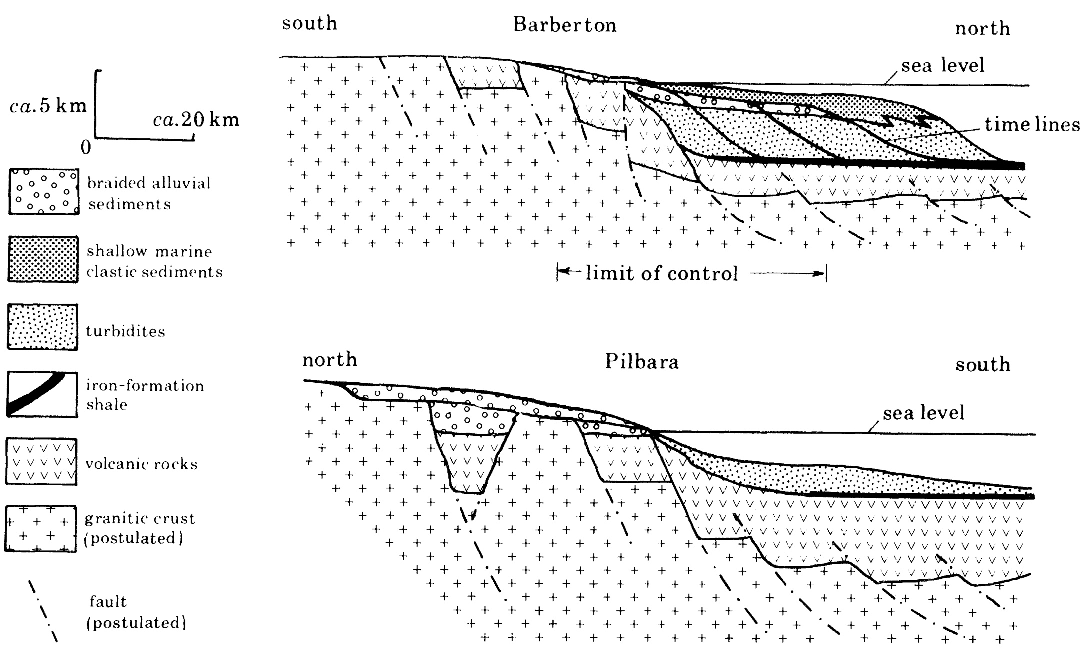

FIGURE 3. Schematic palaeogeographic reconstructions in Barberton and Pilbara greenstone belts, after Eriksson (1980, 1981*b*).

granite–gneiss terrain formed a significant part of the source area for the sediments, and was part of a continental type basement to the greenstone sequences.

Younger *ca.* 2600 Ma greenstone sequences exhibit some similarities to the 3550 Ma Barberton and Pilbara belts, but perhaps not surprisingly exhibit more diversity. It is probable that a range of tectonic–sedimentary environments is preserved and caution should be exercised when trying to formulate models for their tectonic evolution.

Zimbabwe contains younger greenstone belts (2640 ± 140 Ma) (Hamilton *et al.* 1977), which developed over an area of *ca.* 700 km × 200 km on basement containing older sialic and greenstone material. It is possible to correlate the greenstones lithostratigraphically across much of the area (Wilson *et al.* 1978), and in two areas well preserved unconformities are preserved between basal sedimentary units and the older granite–gneiss basement (Bickle *et al.* 1975; Orpen & Wilson 1981). Distinctive features of the stratigraphy are a thin basal clastic sedimentary formation, a very uniform 5–7 km thick sequence of komatiitic and tholeiitic basalts, and an overlying clastic sedimentary sequence (figure 2). Above these widely distributed lower units, the younger greenstone stratigraphy shows regional variation with the development of a major calc-alkaline belt along the *ca.* 700 km northwest margin of the craton. The sedimentology of the Rhodesian greenstone sequences has not been studied in adequate detail, but the preservation of basal unconformities does allow the early subsidence history to be studied, providing information not available in other Archaean areas.

In the Belingwe Greenstone Belt (Bickle *et al.* 1975), the basal sedimentary formation is in unconformable contact with both 3500 Ma gneisses and older greenstone sequences. The unconformity shows only minor relief. The basal sediments pass up from intertidal facies into deeper water, sub-wave base facies, and are capped by a laterally persistent iron formation; subsidence was initiated before the ensuing volcanicity. The overlying volcanic sequence comprises a monotonous 5–7 km sequence of pillowed and massive basaltic flows, with a major komatiite horizon at its base. It contains only minor cross-bedded volcanoclastic sediments and rare cherts. The volcanic rocks are overlain by a second major sedimentary interval of conglomerates derived from the underlying volcanic rocks as well as shallow marine clastic facies, iron-formation and stromatolitic limestones (Martin *et al.* 1980). The volcanic rocks are overlain directly by iron formation in the northwest, and by locally derived conglomerates in the southeast. Both subsidence and localized uplift must have occurred on termination of the volcanicity.

In Belingwe, the sediments are *ca.* 2 km thick, and the top is not preserved. Elsewhere, they pass into a further volcanic association. Facies variation within the Belingwe Greenstone Belt, coupled with uniform north-northwest trending current directions based on cross-bedding measurements, suggests that the currently preserved sequence was not the site of a localized basin, but rather part of a more widespread sedimentary and volcanic blanket.

Canadian greenstone terrains are of similar age to the Zimbabwe younger greenstones (*ca.* 2600 Ma). The greenstones of the Superior Province form a major area of mafic volcanics, felsic volcanics and sediments similar in general features to other greenstone terrains. Like most Archaean areas, evidence for the basement to the greenstone sequences is largely lacking. The stratigraphy is complex, and unconformities occur within the greenstones, with several cycles of volcanicity and sedimentation. A number of unconformities are reviewed by Baragar & McGlynn (1976) but the regional significance of these is still uncertain.

Hyde (1980) has summarized the sedimentology of the Timiskaming Group (figure 2), and points to similarities with sedimentation in other parts of the Superior Province (see, for example, Walker 1978). Sedimentary sequences overlie submarine mafic to felsic volcanic sequences. Two main clastic sedimentary associations are identified: non-marine braided alluvial and deeper water turbidite associations. In contrast to both Barberton and the Pilbara, the marine turbidite association characteristically *overlies* the non-marine association, or is laterally equivalent to it. Hyde suggests non-marine sedimentation on transitory volcanic islands, which ultimately subsided back into the surrounding deeper water turbidite basins. Palaeocurrent directions indicate variably orientated palaeoslopes, analogous to a variety of modern ocean volcanic island environments. The sedimentary provenance is largely volcanic with some granitic detritus derived from subvolcanic plutons. There is no evidence of a granitic provenance as in Barberton or the Pilbara.

The total thickness of sediments with minor intercalated volcanics is between 2 and 3 km in the Timisking area and is similar in several other areas in the Superior Province (Hyde 1980). In all these areas, tops to sequences are not preserved, or terminate with more volcanism. Turbidite sediments predominate near the tops of sequences, implying substantial water depths (at least *ca.* 500 m and possibly were more than 1 km). Henderson (1972, 1975) records *ca.* 5 km of braided alluvial and turbidite deposits in a similar sedimentary association, in the Yellowknife greenstone province.

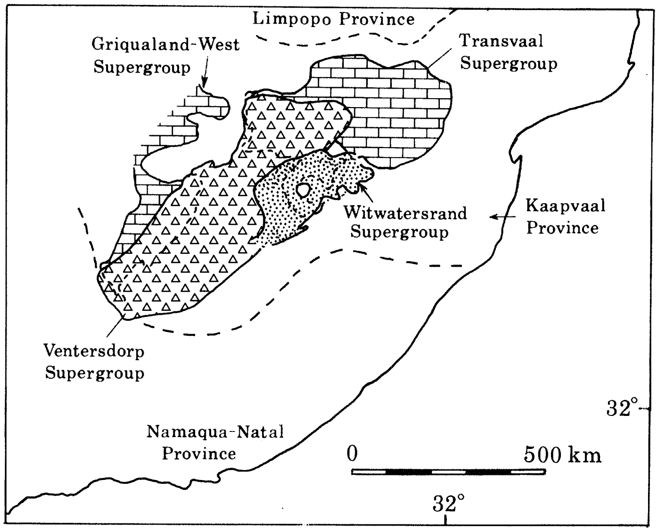

FIGURE 4. Distribution of cratonic basins (3.0–2.0 Ga) in South Africa, stripped to show oldest basin.

Cratonic basins

Major sedimentary basins developed on stabilized Archaean granite–gneiss and greenstone 'cratons' in all the major continents, and are preserved with little deformation. In South Africa, the Kaapvaal province (figure 4) had stabilized by *ca.* 3000 Ma, and four major basins developed in the period 3000–2000 Ma, the Pongola, Witwatersrand, Ventersdorp and Transvaal–Griqualand West supergroups; their distribution on the province is shown in figure 4, and their thicknesses, stratigraphy and available age constraints in figure 5 and table 1.

These four early Precambrian stratigraphic units are separated by province-wide unconformities similar to those documented by Sloss & Speed (1974) for the Phanerozoic history of the United States. The sedimentary basins evolved during *ca.* 200 Ma periods of cratonic downwarping with or without submergence, whereas the unconformities represent periods of cratonic emergence.

(*a*) *Tectonic setting of the early Precambrian cover sequences*

The four early Precambrian cover sequences developed in progressively larger basins and in response to the northward migration of depositional axes; the Pongola Supergroup accumulated in the southeastern corner of the Kaapvaal Province, whereas the Transvaal–Griqualand West Supergroup developed on the northern and western margin of the province. The migration of depositional axes has been related by Hunter (1974) to a complementary migration of loci of granite emplacement north of the corresponding sedimentary basin.

The Pongola, Witwatersrand and Transvaal–Griqualand West supergroups are terrestrial and shallow water cratonic shelf sequences that developed in stable foreland basins fronting uplifted source terrains to the north. In particular, the Pongola Supergroup has been interpreted as the post-orogenic response to closure of a northward-opening passive margin along which the Fig Tree and Moodies groups in the Barberton Mountain Land were deposited (Eriksson 1981 *a*). The Ventersdorp Supergroup accumulated in intracratonic grabens and thus, in contrast to the sheet-like geometry of the aforementioned stratigraphic sequences, displays abundant wedging of lithologies. The middle Ventersdorp sediments in particular thicken dramatically towards fault scarps (fig. 7 in Buck 1980) and, as a result of reactivation of faults, numerous regional surfaces of unconformity occur within the supergroup.

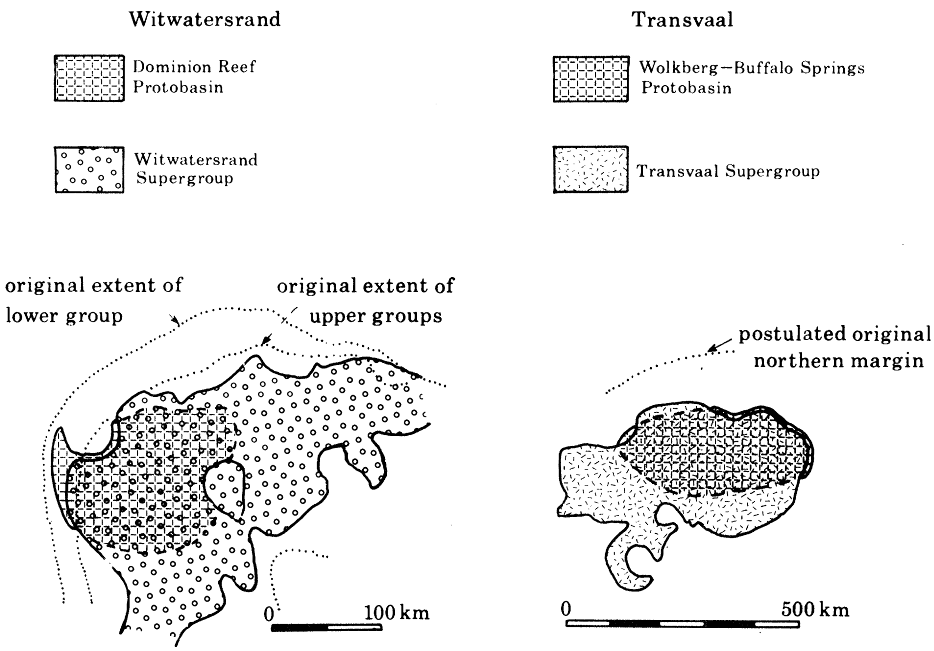

Figure 5. Extent of Witwatersrand and Transvaal proto-basinal and main basin phases, after Pretorius (1976), Button (1973 *b*) and Tyler (1979).

(*b*) *Basin evolution*

(i) *Protobasinal stage.* Of the four early Precambrian stratigraphic sequences occurring on the Kaapvaal Province, the Witwatersrand and Transvaal–Griqualand West Supergroups contain recognizable protobasinal phases. These are represented respectively by the *ca.* 2000 m thick and 150 km wide Dominion Reef Group (figures 5 and 6) and the *ca.* 1600 m thick and 200 km wide Wolkberg–Buffalo Springs Group (figures 5 and 6) (Button 1973 *a*; Tyler 1979). The lithologies of these two protobasinal units are dominated by volcanics and coarse-grained arkosic sediments. The volcanic rocks are mainly continental basalts having some oceanic affinities, although the Dominion Reef and Buffalo Springs volcanics do display bimodal characteristics. Tyler (1979) has documented potassium-rich rhyolites from the Buffalo Springs Group. The Dominion Reef volcanic rocks and the Wolkberg–Buffalo Springs volcanic rocks are predominantly subaerial. The associated arkosic sediments are of braided alluvial origin with palaeocurrent patterns indicating transport both normal and parallel to the axes of the protobasins (Button 1973 *a*; Tyler 1979). The basaltic and bimodal composition of the volcanic rocks in the Dominion Reef and Wolkberg–Buffalo Springs groups favour extensional conditions during the initial stages of basin development. There is, however, no field evidence of graben development associated with either of the protobasins. This may be a function of preservation with the original fault-controlled margins of the protobasins beyond the present outcrop limits of the stratigraphic units.

The Pongola Supergroup does not contain a recognizable protobasin, but begins with arkosic rocks overlying the basal unconformity and passes up into a relatively thick sequence of calc-alkaline volcanic rocks (N. V. Armstrong, personal communication 1979; Kroner *et al.* 1980). The Ventersdorp stratigraphy is dominated by volcanic rocks. The lower Ventersdorp volcanic rocks are conformable on sediments of the Witwatersrand Supergroup, are laterally continuous and are considered to be the the terminal phase of the Witwatersrand Supergroup.

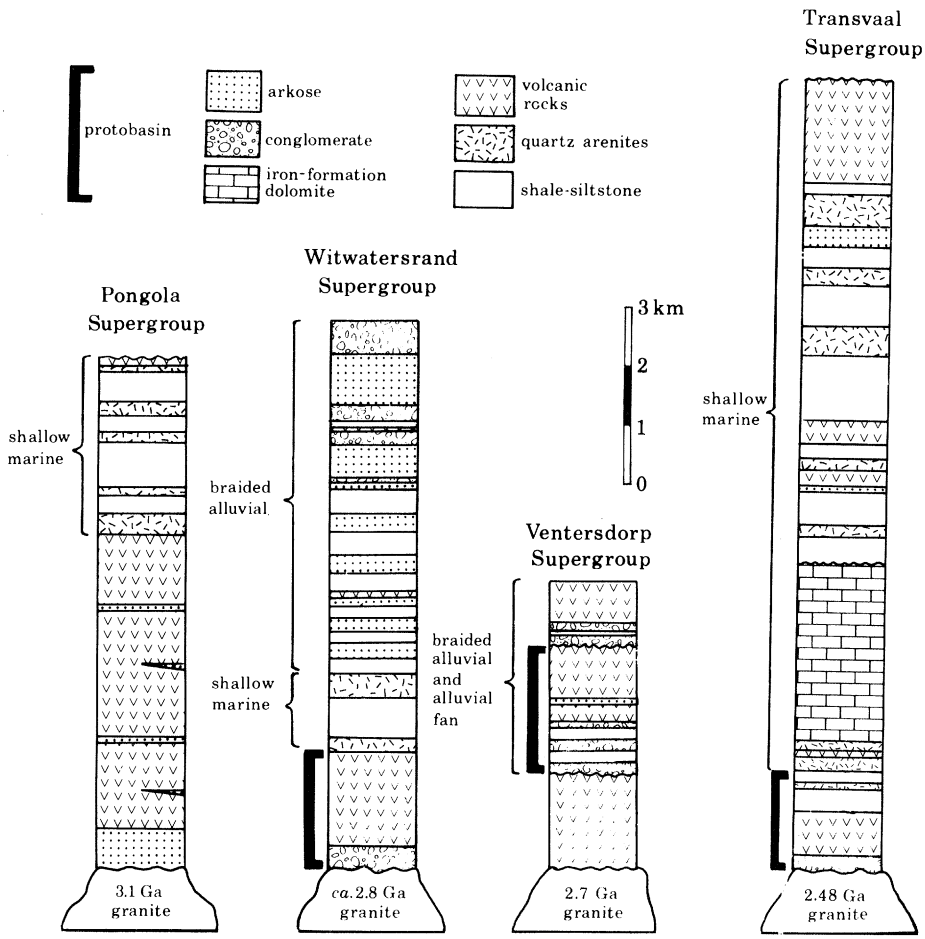

FIGURE 6. Stratigraphy of Pongola, Witwatersrand, Ventersdorp and Transvaal basins (references in table 1).

Graben formation occurred above these volcanics (figure 7), and may represent the initial subsidence phase of the Ventersdorp Supergroup. Alluvial fan braided alluvial and lacustrine sediments interbedded with volcanic horizons developed in the grabens.

(ii) *Subsidence stage.* The protobasinal to subsidence stage transition is well displayed in the Witwatersrand and Transvaal Supergroups. Although unconformable along the margins of the protobasins, the transition becomes conformable and gradational towards the centre of the basins implying limited erosion during the transition (figure 8). Rather, subsidence occurred over a much greater area to produce depositional basins with minimum widths of 200 km for the Witwatersrand Supergroup and 300 km for the Transvaal Supergroup (figure 5). On the basis of palaeocurrent and facies patterns, Button (1973*b*) and Eriksson *et al.* (1976)

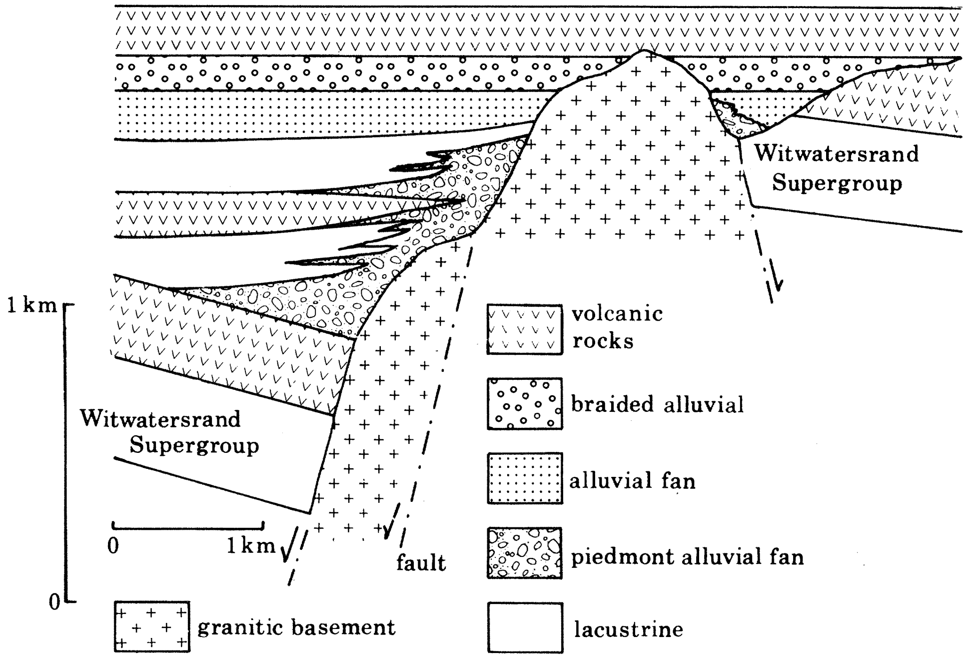

FIGURE 7. Schematic section through Ventersdorp Supergroup, after Buck (1980).

proposed that the post-Wolkberg Transvaal depository may have been as much as 600 km in width.

In both the Witwatersrand and Transvaal supergroups, the subsidence stage began with a marine transgression, resulting in the drowning of the Kaapvaal Province. The earliest sediments in both these basins are considered to have accumulated under conditions of tidal reworking (Button 1973*b*; Eriksson *et al.* 1981). In the Witwatersrand Supergroup, the lower tide-dominated sequence of quartz arenites and shales passes upwards into conglomerates and arkoses of braided alluvial origin (figure 6). Shallow marine conditions prevailed almost throughout the history of the Transvaal–Griqualand West Supergroup with the 10 km thick sequence of chemical and terrigenous clastic sediments (figure 6) accumulating in various tidal environments (Eriksson *et al.* 1976; Button & Vos 1977). The shallow marine sediments in both supergroups are characteristically arranged in progradational, upward-coarsening sequences, indicating an excess of sediment supply over basin subsidence. The sequences are terminated along surfaces of transgression, which developed upon cessation of sediment supply.

Volcanic horizons are present within both the Witwatersrand and Transvaal supergroups, and compose *ca.* 7% of the latter (figure 6). Tholeiitic basalts are again predominant but the uppermost volcanic horizon in the Transvaal Supergroup does contain acidic phases (Button 1973*b*). Superimposed on the overall subsidence history of the Witwatersrand and Transvaal–Griqualand West supergroups were periods of emergence during which internal surfaces of unconformity developed. Particularly striking is the regional unconformity at the top of the chemical sedimentary unit in the Transvaal Supergroup. Volcanicity continued throughout the Ventersdorp Supergroup, and it is not possible to separate thermal subsidence from initial subsidence unambiguously. The relatively thin, but laterally more continuous, upper fluvial sediments (figure 7) may represent thermal subsidence, but these are terminated by further continental basalts.

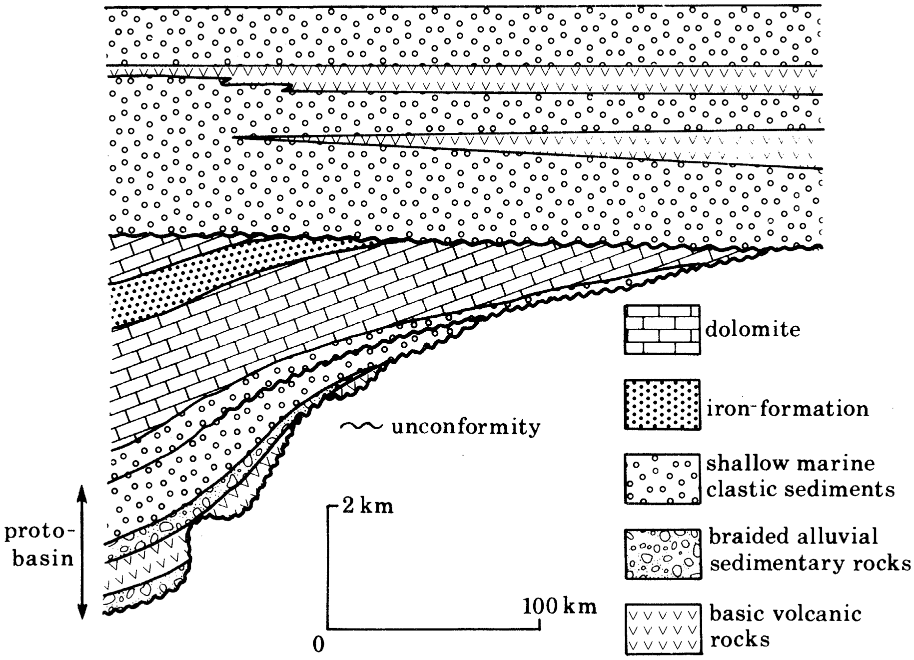

FIGURE 8. Schematic section across Transvaal Supergroup, after Button (1976).

TECTONIC IMPLICATIONS

Distinction of initial and thermal subsidence phases

Recognition and separation of the initial from the thermally controlled subsidence is the basis for our interpretation of early Precambrian sedimentary basins. In all the basins discussed above, the volcanic rocks compose a major interval low in the sequence, and in two examples (Witwatersrand and Transvaal supergroups) these volcanics are confined to a spacially restricted protobasin. There are a number of reasons to associate the volcanicity with the initial phase of subsidence, and we shall argue that the termination of volcanicity in the majority of the basins discussed above marks the cessation of the initial subsidence phase.

Modern volcanic rocks are invariably erupted in tensional environments, and there are physical constraints on the transport of magmas that require this (Weertman 1971). Volcanic intervals may therefore provide *a priori* evidence for extension (although it should be noted that the volcanic and sedimentary facies do not always indicate graben development indicative of crustal extensions). Where volcanic horizons occur higher in the sedimentary sequence, these may indicate renewed tectonic activity, and models based on one initial instantaneous tectonic event may not be appropriate. Transport of magmas from mantle depths may also provide sufficient heat to thin the lithosphere and thus subsidence associated with the volcanic event would be followed by subsidence related to thermal relaxation of the lithosphere.

The most compelling reason to associate the volcanic-dominated intervals with initial subsidence is that, to accommodate up to 7 km of volcanic rocks, substantial and rapid subsidence must have occurred. Subsidence in greenstone belts cannot be due to loading an excess thickness of volcanic rocks, as the volcanic rocks were either erupted in shallow water as evidenced by intercalated cherty sediments (Barberton and Pilbara) or are immediately underlain and overlain by shallow marine sediments (Zimbabwe). In the Witwatersrand, Transvaal–Griqualand West and in the Ventersdorp supergroups, volcanic rocks are mainly subaerial, although in the Witwatersrand and the Transvaal shallow marine sediments overlie the volcanics requiring net subsidence during the initial phase. This is also inconsistent with subsidence being only due to loading by the sedimentary–volcanic pile. The small proportion of interbedded sediment within the volcanic intervals, relatively minor horizons of reworked volcanoclastic material and lack of weathered or soil horizons within the subaerial volcanic intervals, all attest to the rapidity of the volcanism and thus of the initial subsidence phase. The shallow marine setting of greenstone volcanic intervals allows accurate estimates of total subsidence during the initial phase. The subaerial to shallow marine transition across the initial subsidence phase in the Witwatersrand and Transvaal may have required more subsidence than implied from sedimentary–volcanic thicknesses alone, although deposition of major alluvial deposits more than a few hundred metres above sea level is uncommon today.

Marginal unconformities passing into continuous depositional sequences in basin centres may mark the termination of the initial subsidence phase. In all of the greenstone-associated basins, the initial volcanic phase is terminated by the uplift and erosion of parts of the volcanic sequence. In the protobasin underlying the Transvaal, shallow marine sediments *ca.* 400 m thick overlie the volcanics, but along the margins of the basin, both these and the volcanics were eroded. This uplift is consistent with volcanicity being associated with initial subsidence, as Beaumont (1978) shows that marginal recovery over *ca.* 10 Ma would reduce the area of depression caused by the elastic response to a load on the lithosphere. The only constraint is that the viscous recovery must be more rapid than the ensuing thermal subsidence, which resulted in deposition over a larger area than that occupied by the protobasin.

Additional evidence to support this assertion of volcanic-dominated initial subsidence might be expected from two other sources. First, the tectonic setting of the volcanics would be expected to provide evidence of the tectonic mechanism for initially thinning or loading the crust and, secondly, subsidence following the initial subsidence should take place at an exponentially decreasing rate. In the early Precambrian and lacking precise geochronology, estimates of sedimentation rates are virtually impossible. Only in the 3550 Ma Barberton and Pilbara greenstone areas is there evidence of rapid thermal subsidence in the form of deeper water iron-formation, shale and then major turbidite sequences deposited on shallow-water volcanic rocks. In both these areas, there is evidence of prograding sedimentary sequences, best displayed in Barberton by the transition from turbidite to braided alluvial and shallow marine deposits. In the other basins, shallow marine and alluvial sediments fill the entire sequence; presumably, sediment supply was sufficient to keep the basin filled, and excess sediment was transported across the basin.

In some basins, there is evidence of active faulting and graben formation during the initial subsidence phase consistent with an extensional origin for the basins. In both Barberton and the Pilbara, Eriksson (1981 *a*) proposes steep-rift type continental margins dominated by major fault scarps to account for the abrupt turbidite to braided alluvial transition (figure 3). However, there is less evidence for active faulting during the volcanicity; volcanoclastic sediments are of minor importance except where directly related to felsic volcanic centres, and the lack of granitic detritus within the volcanic sequences implies that basement was not exposed during this phase (Lowe 1980). Similarly, in Zimbabwe, both the thin basal sediments and the overlying volcanic stratigraphy are laterally continuous.

In the Witwatersrand and Transvaal protobasins (the Dominion Reef and Wolkberg–Buffalo Springs groups), there is no evidence for fault-controlled volcanicity or sedimentation. Much of the Dominion Reef Group is hidden under the deeper undrilled parts of the Witwatersrand basin, but much younger (late Karoo) uplift has exposed a cross-section through the central parts of the Transvaal Basin. Button (1973 *a*) records considerable topographic control on early deposition of the Wolkberg Group, but no syn-depositional faulting (figure 8). It is difficult to reconcile this with an extensional origin for the basin. The rapid lateral facies variation and the thickness changes within the Ventersdorp (Buck 1980) (figure 7) demonstrates the ease with which fault control on sedimentation may be recognized.

Implications for lithosphere thickness

The discussion above shows that the sedimentary history of a number of early Precambrian basins can be related to a relatively rapid initial subsidence phase followed by more widespread, mainly volcanic-free, sedimentation. Sedimentation in the initial phase in all the basins studied is either shallow marine or fluvial, allowing reasonable estimates of the initial subsidence to be made. In several of the basins, the second subsidence phase also terminates with shallow marine or alluvial sediments. On the basis of modern analogues, we shall assume that the second phase can be related to thermal relaxation of the lithosphere and that the initial phase must reflect the prime tectonic control on basin subsidence, either crustal extension or some other crustal loading process. The total basin subsidence, initial plus thermal, depends only on the ultimate cause, either the amount of crustal thinning or the magnitude of crustal loading. The thermal subsidence relates directly to the increase in lithosphere thickness. We have estimated sedimentary thickness, water depths, and densities for the basins discussed above (table 1), and have calculated the equivalent crustal thinning for probable crustal thickness from the formula

$$1/\beta = 1-\frac{S(\rho_s-\rho_o)}{t_c(\rho_c-\rho_o)},$$

where β is the amount of crustal extension, ρ_o is mantle relative density (3.33), ρ_c is the relative density of the continental crust (2.8), ρ_s is the relative average density of sediment–volcanic infill and S is the total subsidence.

We have then calculated lithosphere thickness from the thermal subsidence on the assumption that lithosphere thinning is inversely proportional to crustal extension. Lithosphere thickness, a, is given by

$$a = \frac{2S_t(\rho_o-\rho_s)}{\alpha T_1(1-1/\beta)\rho_o},$$

where S_t is the thermal subsidence, T_1 the temperature of the base of the lithosphere (1333 °C) and α the coefficient of thermal expansion (3.28×10^{-5} K^{-1}). This calculation assumes a linear thermal gradient through the lithosphere, crust and sediments, and ignores heat production, but given uncertainty on α (*ca.* $\pm 1 \times 10^{-5}$) and T_1 ($\pm$ *ca.* 100 °C) a more sophisticated treatment seems unjustified.

Table 2. Stretching model parameters for selected early Precambrian basins

(Note that increase in lithosphere thickness gives minimum lithosphere thickness if relation between initial equivalent crustal thinning and initial lithosphere thinning is not valid.)

	thickness of continental crust† (if initially at sea level)/km	extension† (β)	equilibrium† lithosphere thickness km	increase in lithosphere thickness for thermal subsidence km
Greenstone belts, 3.5 and 2.6 Ga	13.7 (minimum)	∞	59	59
total subsidence 12.5 km, $\bar{l}$ = 2.75	20	3.2	86	
thermal subsidence 5.5 km, $\bar{l}$ = 2.55	30	1.8	133	
	40	1.5	177	
Pongola	10.9 (minimum)	∞	32	32
total subsidence 8.5 km, $\bar{l}$ = 2.65	20	2.2	59	
thermal subsidence 3 km, $\bar{l}$ = 2.55	30	1.6	86	
	40	1.4	112	
Witwatersrand, 2.7 Ga	12.2 (minimum)	∞	75	75
total subsidence 9 km, $\bar{l}$ = 2.61	20	2.6	121	
thermal subsidence 7 km, $\bar{l}$ = 2.55	30‡	1.7	182	
	40‡	1.4	262	
Ventersdorp, 2.64 Ga	4.0 (minimum)	∞	5.4	5.4
total subsidence 3.5 km, $\bar{l}$ = 2.72	20	1.25	27	
thermal subsidence 0.5 km, $\bar{l}$ = 2.55	30‡	1.15	41	
	40‡	1.11	54	
Transvaal, 2.2 Ga	15.2 (minimum)	∞	92	92
total subsidence 12.0 km, $\bar{l}$ = 2.66	20	4.1	121	
thermal subsidence 10.0 km $\bar{l}$ = 2.66	30‡	2.0	184	
	40‡	1.6	245	

† Assumes initial subsidence is a result of thinning crust and lithosphere by stretching only.
‡ Likely estimates for initial crustal thickness.

The lithosphere thickness estimated depends directly on the assumption that the initial thinning was proportional to the effective thinning of the crust. As discussed above, even if an extensional model is appropriate, transport of heat by magmas may over-thin the lithosphere. The only reliable figure is the minimum lithosphere thickness appropriate to infinite extension in table 2, which gives the amount by which lithosphere thickness must increase to cause the observed thermal subsidence.

The results of these calculations are given in table 2. The Ventersdorp Supergroup, although possessing a well developed volcanic–sedimentary initial phase, apparently developed only a very limited phase of thermal subsidence. Calculated minimum lithosphere thicknesses range from *ca.* 60 km in the greenstone areas to *ca.* 90 km for the Transvaal. In Zimbabwe, Witwatersrand and the Transvaal, there is evidence for a significant thickness of continental crust preserved beneath the basins, and for the crust (*ca.* 30 km in South Africa) to be preserved, lithosphere thickness must not have been reduced to less than *ca.* 40 km (including the crust). Realistic minimum equilibrium lithosphere thicknesses must have ranged from *ca.* 75 km in greenstone terrains to 110–130 km under the cratonic basins in South Africa at 2.6–2.2 Ga. For the two South African basins, the present crustal thickness of *ca.* 36 km (Gane *et al.* 1956) enables constraints to be placed on the amount of extension, if this was the fundamental cause. Factors of between 1.4 and 1.7 for Witwatersrand and 1.6–2.0 for the Transvaal would require *ca.* 200 km thick lithospheres. The total time required for thermal relaxation of such thick lithosphere (*ca.* 500 Ma) is far greater than the *ca.* 200 Ma available and suggests that simple extension is incapable of explaining the thermal subsidence alone. If rifting was the cause, some mechanism must have caused over-thinning of the lithosphere, as suggested by Steckler & Watts (1980).

The validity of these conclusions depends on the correct identification of the initiation of thermal subsidence and, as noted by McKenzie *et al.* (1980), precise estimates of the duration of subsidence will provide the only possible independent verification of these figures. Uncertainty in densities and the coefficient of thermal expansion also cause uncertainties on the estimates of the order of $\pm 30\,\%$.

Conclusions

Early Precambrian sedimentary basins occur in a number of different tectonic settings, including greenstone belts, cratonic shelf and cratonic rift environments. All the basins contain major volcanic horizons low in the sedimentary sequences. There are a number of reasons why these volcanics are thought to be associated with initial subsidence of the basins: (1) volcanicity was restricted to protobasins in some examples; (2) volcanicity was followed by marginal uplift and erosion, but subsidence continued in basin centres; (3) thick volcanic sequences erupted entirely in shallow marine conditions imply significant and rapid subsidence; (4) there are theoretical reasons why volcanicity might be associated with basin formation: eruption of volcanics requires a tensional environment and transport of heat by volcanics provides a mechanism to thin the lithosphere, allowing the subsequent protracted thermal subsidence. We postulate that the initial volcanic-dominated subsidence was a direct result of the tectonic mechanism, which thinned or loaded crust, permitting permanent preservation of the basin. Sedimentary facies relations in some greenstone terrains and in the Ventersdorp cratonic rift sequence are consistent with an extensional tectonic cause for basin formation in these areas. Cratonic shelf basins such as the Witwatersrand and the Transvaal, although containing volcanic-dominated initial phases preserved in protobasins, did not apparently develop above a rifted crust, and some other mechanism must be sought

Distinction of the initial subsidence phase from the subsequent 'thermal' subsidence phase related to thermal relaxation of a thinned lithosphere enables constraints to be put on minimum lithosphere thicknesses. Such estimates imply equilibrium lithosphere thicknesses greater than *ca.* 75 km in greenstone terrains at 3.5 Ga, *ca.* 75 km in greenstone terrains at *ca.* 2.6 Ga, greater than *ca.* 110 km in cratonic shelf basins at 2.6 Ga, and greater than *ca.* 130 km in cratonic shelf basins at 2.2 Ga. If our distinction of initial subsidence from thermal subsidence is correct, then the lithosphere in the early Precambrian was comparable in thickness with that of the present day.

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Discussion

D. G. Roberts. How do the authors obtain the lithospheric thickness in the Archaean? The basin subsidence was presumably produced partly by the load of sediments infilling an existing basin and perhaps partly by thermal contraction. How can one distinguish between these two mechanisms?

M. J. Bickle. We assumed a sediment density and corrected for the sediment load in the usual way to obtain the subsidence history the basin would have had if it had been filled by water only. The principal error is the uncertainty in the sediment density at the time of loading.

D. Hastings. I tend to agree with the authors' hypothesis that at least some early Precambrian sedimentary basins (greenstone belts) formed as a result of rifting. Many scientists, however, argue that the tightly folded rocks of many such features require a strongly compressive environment. Are the authors' arguments for a tensile origin of the belts compatible with theirs?

Incidentally, do the authors see their tensile environment in terms of parallels to Cainozoic (postulated) early stages of ridge-like extension which failed to develop very far, to postulated back-arc environments (do they see any examples at all of early Precambrian subduction?), or another possible model?

M. J. Bickle. The intense folding that commonly deforms the rocks of greenstone belts has been produced by later events. In the Pilbara the sedimentary volcanic sequence formed at 3.55 ± 0.02 Ga. The next event was the intrusion of granodiorite bodies at about 3.35. Ga. The original sediments were still essentially undeformed when these rocks were intruded. The sediments were then strongly folded, in places even recumbantly folded. We did not use thicknesses obtained in such places to estimate the subsidence. We therefore do not believe basin formation and subsequent tight folding are directly related.

B. Chadwick. I wish to describe some results of recent field investigations in collaboration with Dr M. Ramakrishnan and Mr M. N. Viswanatha, Geological Survey of India, in the Chitradurga region (76° 20′ E; 14° 10′ N) in the Karnataka craton of southern India which appear to support certain aspects of the models for the evolution of late Archaean basins proposed by Dr Bickle & Dr Eriksson.

The late Archaean Dharwar Supergroup (*ca.* 2600 Ma) in southern Karnataka occurs in a series of belts and irregular basins resting unconformably or with tectonic contact on a basement of tonalitic–granitic Peninsular Gneiss (*ca.* 3000 Ma) which contains enclaves and locally extensive tracts of an older supracrustal association called the Sargur Group or Sargur high-grade schists. Detailed descriptions are provided by Swami Nath & Ramakrishnan (1981).

The evolution of the western part of the late Archaean basin in the Chitradurga region is shown schematically in figure 1. The succession in the Dharwar Supergroup begins with basal

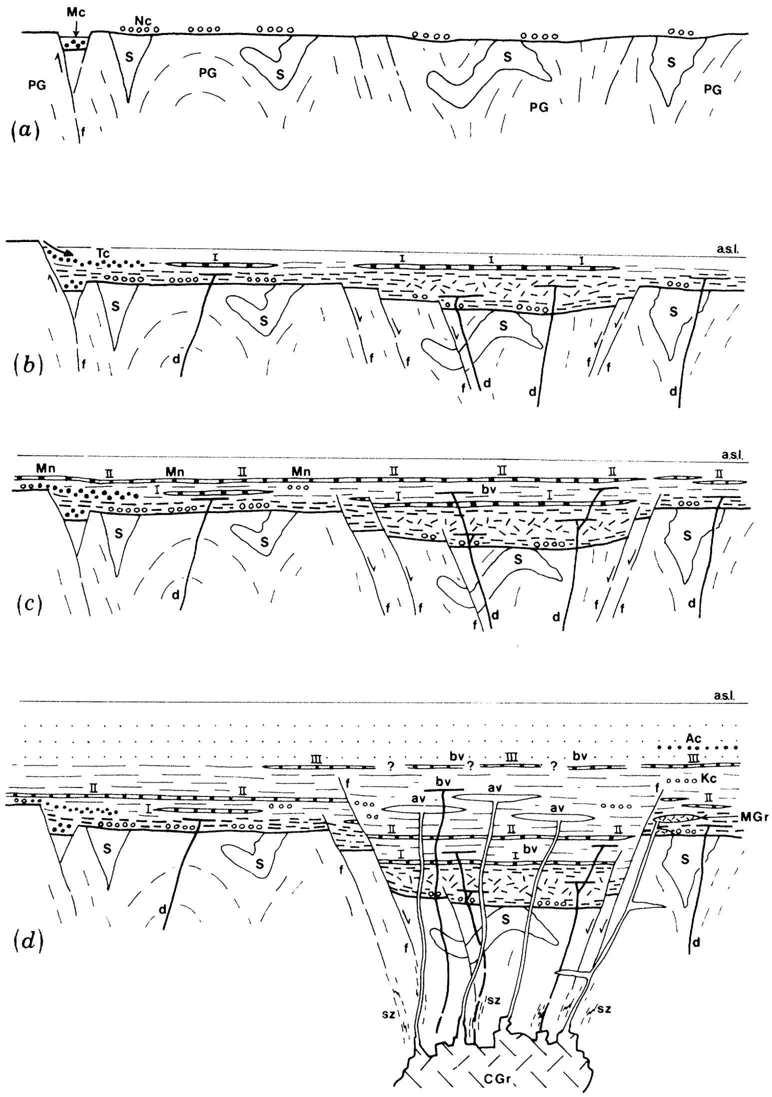

FIGURE 1. Schematic representation of the stratigraphic evolution of the Dharwar Supergroup in the Chitradurga region, Karnataka. S, Sargur Group; PG, Peninsular Gneiss; Mc, Mayakonda polymict conglomerate; Nc, Neralekatte quartz conglomerate; Tc, Talya polymict conglomerate; I–III, banded iron formations; bv, basic volcanic rocks; av, acid volcanic rocks; Kc, Kurmeredikere conglomerate; Ac, Aimangala polymict conglomerate. Thick dashed lines, basic volcanic rocks; thin lines, siliceous phyllites and cherts; dotted ornament, greywacké-turbidites, a.s.l. approximate sea level; sz, f, schematic representation of shear zones and faults active during the depositional and volcanic phases. Estimated width of section 25–30 km.

quartz pebble conglomerates and quartzites (figure 1*a*) which appear to have been deposited after peneplanation and mature weathering of the basement. Basic volcanism and sporadic introduction of cross-bedded, clean-washed sands characterize the initial platformal environment that we believe to have covered much of the Karnataka craton at this time. Thicker basic volcanic rocks in a more internal part of the basin (figure 1*b*) indicate greater subsidence of the basement, which we relate to rifting and graben formation. Polymict conglomerates (Tc, figure 1*b*) containing clasts of basement and Dharwar cover sediment and volcanic rocks formed near the margin of the basin. Continued subsidence accommodated siliceous phyllites, cherts and banded iron formations (figure 1*c*). Certain iron formations are Mn-bearing in the marginal parts of the basin. Greater rates of subsidence in the more internal parts appear to have been maintained throughout the stratigraphic evolution of the belt to accommodate sequences of basic and acid volcanic rocks (figure 1*d*). Progressive deepening of the whole basin took place to accommodate the youngest rocks, namely greywacké-turbidites and local polymict conglomerates (Ac, figure 1*d*). Estimated total stratigraphic thickness is at least 10 km in the deeper parts of the basin.

We suggest that a genetic relation may exist between the Chitradurga granite (CGr, figure 1*d*) and the acid volcanic rocks, a relation that appears to be supported by isotope data (P. N. Taylor, personal communication 1981). The Chitradurga granite (*ca.* 2600 Ma) rose through the cover during the deformation that followed the depositional and volcanic phases. The granite now appears as a post-tectonic intrusion. Deformation led to the formation of a series of mainly upright anticlines and synclines and related LS fabrics with extreme variations in the plunge of coaxial fold axes and L fabrics. Low-grade metamorphism outlasted the deformation. The deformation followed soon after the depositional and volcanic phases.

The stratigraphy of the Dharwar Supergroup in the Chitradurga region is consistent with the proposal by Bickle & Eriksson that late Archaean basins began with a protobasin phase dominated by basic volcanism. Subsequent stages of their model, including erosion and uplift of part of the basin, are also evident in the occurrence of various polymict conglomerates containing clasts of the Dharwar cover and its basement gneisses. However, the volcanic phase in the Chitradurga belt does not appear to have been confined to the protobasin phase because in the more internal part of the basin, basic and acid volcanism continued relatively late in the depositional phase. The thicker development of volcanic rocks appears to be directly related to greater subsidence. Rifting appears to have played an important role in the evolution of the Chitradurga belt. Deformation and granite emplacement also appear to have followed soon after the close of the depositional and volcanic evolution of the basin. This subsequent tectonic and igneous activity may have been closely related to the sedimentary and volcanic events.

Patterns of events in late Archaean belts may vary. For example, the younger greenstone belts of Zimbabwe described by Bickle and Eriksson appear to have much thicker basalt–komatiite associations compared with the Dharwar Supergroup of Karnataka. The differences in type and degree of volcanism and variations in types of associated sedimentary rocks suggest that while the proposals by Bickle & Eriksson will provide a valuable stimulus for research into the evolution of Archaean sedimentary basins, much detailed investigation of Archaean supracrustal associations remains to be done to establish the general validity of their models.

Reference

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A. M. Clarke. Have the authors seen any evidence in the Precambrian for a clear separation between sediments laid down during the rifting phase and those deposited during the thermal and flexural phases of basin formation? Have they found any examples of the superposition of a later on a younger basin, as Kerr has described during the development of the Sverdrup basin?

K. A. Eriksson. The three basins in South Africa cover a time interval of 600 Ma, and in places are superimposed on each other to produce a sediment thickness of 15 km. Each basin shows an area of limited extent, the protobasin, of initial subsidence followed by subsidence over a larger region. These protobasins are, however, aerially separated.

P. A. Ziegler. The stratigraphic record of the basins discussed by the authors indicates that volcanic activity persisted, albeit with intensity variations, throughout their evolution. As volcanic activity cannot be reconciled with the lithospheric cooling and contraction model of inactive rifts and continental margins it is questioned whether these basins had ever entered a clear post rifting stage.

Could Dr Bickle provide further information about the late volcanism in these basins such as chemistry, mode of emplacement of volcanics and possible tectonic causes of volcanism?

M. J. Bickle. Only about 7% of the sequence consists of volcanic material. If the thermal structure of the lithosphere is to be strongly affected by volcanism, volumes of magma comparable with that of the continental crust must be intruded. Unless there was further crustal thinning even such large amounts of volcanism would only delay the onset of the thermal subsidence phase.

K. A. Eriksson. The volcanics are separated by well developed soil horizons, which perhaps could have resulted from pauses in the thermal subsidence produced by the heat from the magmas.

M. F. Osmaston. Could the Precambrian basins have been produced by doming, followed by erosion of the dome, as has been argued happened in the North Sea during the Mesozoic?

M. J. Bickle. The existence of superimposed basins suggests that uplift and erosion did not play an important part in basin formation.

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