Formation of an Archaean continent

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About 30% of the Earth is covered by continents, but only about 10 small kernels of these continents—known as Archaean cratons—are continental fragments formed before 2.5 Gyr ago. The Kaapvaal craton of South Africa, which formed and stabilized between 3.7 and 2.7 Gyr ago, is one of the oldest reasonably sized examples of these continental fragments. It consists of a mosaic of subdomains that have been welded together by processes similar to those of modern-day plate tectonics. The earliest subdomains may have owed their origin to the onset of efficient recycling from the Earth's hydrosphere into the mantle.

SOME of the most fundamental problems concerning the Earth's evolution relate to the first emergence of the continental lithosphere and to the associated tectonic and thermal processes then operating. There is as yet no consensus as to when and how the earliest continental regions were formed, and how much and at what rates continental lithosphere was extracted from the mantle in the Archaean[1]. Although no continental rocks older than about 4.0 Gyr have been found[2], evidence from minerals (zircons) indicates that continents in some form may have existed as far back as 4.2 Gyr (ref. 3). The lack of preservation of continental regions older than 4.0 Gyr may indicate that earlier continents were unstable and were recycled back into the mantle, perhaps in response to intense extraterrestrial bombardment[4]. It is also possible that before 4.0 Gyr, continents on Earth were very rare. If so, we need to explain the cause of the sudden increase in continent formation after 4.0 Gyr, and how the Earth evolved from a state in which it was completely covered by oceans to its present configuration.

It is generally accepted that continents at present grow through a combination of subduction-related additions of new materials from the mantle and amalgamation of pre-existing lithospheric fragments of various sizes, often referred to as terrane accretion[5]. The growth of continents, however, is not synonymous with the concept of continental lithosphere growth: the latter refers to the net addition of new mantle-derived material to the continents, and thus takes into consideration accretion as well as recycling of continental material back into the mantle. Continental lithosphere growth remains a controversial topic, in part because there is disagreement about the rate of continental recycling, which has almost certainly varied greatly since the Archaean[1]. Meaningful analysis of these processes is essential if we are to understand the formation of the earliest continental fragments.

The crust of the earliest continental fragments is predominantly a variable mixture of plutonic rocks with calc-alkaline-like chemistry and supracrustal rocks such as sediments and volcanic rocks, which are often green in appearance. The former are traditionally referred to as 'granites', and the latter as 'greenstones'; together, these rock sequences make up the 'granite–greenstone terrains' of the Archaean cratons.

Many Archaean cratons consist predominantly of granite–greenstone sequences formed in the late Archaean (2.5–3.0 Gyr ago). Those cratonic areas that have an older history have generally been extensively deformed and metamorphosed, and thus lost much of their original character. The Kaapvaal craton in southern Africa and the Pilbara craton of northwestern Australia are the only regions that have retained a substantial portion of relatively pristine mid-Archaean rocks (3.0–4.0 Gyr ago). Because of the enormous accessible mineral wealth of the Kaapvaal craton, this region (Fig. 1) has probably been more extensively investigated than any other Archaean craton.

The Archaean history of the Kaapvaal craton (or simply the craton) spans about 1 Gyr and can be conveniently subdivided into two periods, each of about the same length as the Phanerozoic period (see Table 1). The first period, ~3.7–3.1 Gyr ago, records the initial separation of the continental lithosphere of the craton from the mantle. This period of continental growth resembles plate tectonic processes occurring in modern-day ocean basins, but the difference is that in the mid-Archaean, these oceanic processes seem to have occurred in shallower water depths. The oldest oceanic rock sequences that have retained evidence for shallow submarine processes, like those

TABLE 1 First 1 Gyr of formation of the Kaapvaal continent

		Time (Gyr BP)
Kaapvaal shield formation*	500 million years of intraoceanic tectonics	
	First regional emergence of Kaapvaal continental lithosphere through intraoceanic obduction processes	3.7–3.2
	Western Pacific-style intraoceanic amalgamation of oceanic terranes	3.3–3.2
	Widespread within-shield melting, granite formation and chemical differentiation in the upper lithosphere	3.2–3.1
	Kaapvaal shield stabilized by	3.1
Kaapvaal craton formation*	500 million years of inter- and intra-continental tectonics	
	Regional extension of the Kaapvaal shield; passive continental margin and rift basins	3.1–2.9
	Cordillera-type accretion of composite terranes along north and west margin of Kaapvaal shield; intermontane and foreland basins	2.7–2.9
	Alpine or himalayan-type continent–continent collision between the Kaapvaal and Zimbabwean shields	2.68 ± 0.05
	Foreland extensional and impact-generated rifts	2.7–2.6
	Kaapvaal craton stabilized by	2.6

* Geologists refer to that portion of a continent which has attained stability and which has been little deformed for a prolonged period as a craton. The inner portion of a craton is referred to as a shield. Shields themselves are initially unstable and subjected to extensive deformation and internal chemical differentiation before cratonic rigidity is attained.

occurring at mid-ocean ridges, are found in the Barberton and adjacent granite-greenstone terrains of the southeastern regions of the craton, and within the Pietersburg greenstone terrain along the northern margin of the craton (see Fig. 1).

The second period of the craton's Archaean history (3.1-2.6 Gyr) records intra-continental and continental-margin processes: continental growth during this period occurred mainly through a combination of tectonic accretion of crustal fragments and subduction-related igneous processes in much the same way as has been documented along the margins of the Pacific and Tethys oceans since the Mesozoic. Examples are best observed in the northern parts of the craton, where the 'continental-shelf' terrain of the Central Limpopo mobile belt (Fig. 1*a*, inset A) has been tectonically juxtaposed between the granite-greenstone terrains of the Kaapvaal and Zimbabwean cratons (Fig. 1); and in the western margin of the Kaapvaal craton, which consists largely of late Archaean amalgamated granite-greenstone fragments. Evidence for tectonic processes similar both in style and time to those along the northern and western craton has also been found in the older central and southern parts of the craton, where these tectonic processes are closely related to formation of major late Archaean sedimentary basins and to the final stabilization of the craton.

We argue here that the history of the Kaapvaal craton can be understood in terms of processes similar to those that occur today at mid-ocean ridges, ocean subduction zones and continental margins. During the first period of continent formation, intraoceanic processes dominated, resulting in continental nucleii (shields) which may have been similar to the oceanic plateaux now present in the Western Pacific. We argue, moreover, that the first separation of this early continental lithosphere could have occurred only when the mean elevation of mid-ocean ridges had sunk below sea level. During the second period, the continental nucleii amalgamated by means of collisional processes which, together with internal chemical differentiation, created the first stable continental kernels (cratons).

We now give an overview of the Kaapvaal craton to provide a framework within which to discuss its formation.

Regional framework of Kaapvaal craton

The craton covers about 1.2×10^6 km^2 (Fig. 1). Geological mapping and geochronology have largely defined the outer limits at surface, but geophysics and image processing have further helped in locating the craton boundaries[6,7]. To the north, the craton is bounded by the Zimbabwean craton; to the west and south, it is bordered by Proterozoic mobile belts; and to the east by the Lebombo monocline of Jurassic volcanics associated with the break-up of Gondwana (Fig. 1*a*). In depth, the craton has also been reasonably well defined by seismologists and mantle petrologists. Waveform inversion studies have demonstrated that there is a strong correlation between the surface expression of the craton and velocity variations from crust to upper mantle[7,8]. Crustal refraction observations in the Proterozoic mobile belts contrast with the craton refraction. Results suggest that the crust of the Namaqua Proterozoic belt has a substantial (30 km) component of intermediate-velocity rocks (6.6-6.9 km s^{-1}) and a depth to Moho of about 42 km. In contrast, the intermediate-velocity rocks in the Kaapvaal craton are represented by only about 15 km of the lower crust, and the depth to the Moho is 37 km (ref. 8).

The upper layers of the craton have low S-wave velocities, with a mean of 4.3 km s^{-1} in the crust and upper mantle down to 80 km depth (about 8% slower than global averages); this is followed by a positive velocity gradient and a higher velocity of about 4.6 km s^{-1} between 110 and 220 km depth[7]. The velocity inversions show that velocities beneath the craton in the depth range 50-350 km are substantially greater than the global average; from this data it can be concluded that the maximum base of the craton is at a depth of about 350 km (ref. 7). This is consistent with the idea that cratons are associated with high-velocity anomalies world-wide extending to 300-400 km in depth; Jordan[9] termed this the continental tectosphere.

The craton is characterized by low heat flow (~45 mW m^{-2}) and the occurrence of diamondiferous kimberlites, whereas the younger surrounding mobile belts have relatively high heat flow (~80 mW m^{-2}) and barren kimberlites, and thus have lithosphere thicknesses less than 120 km (ref. 10). Geotherms calculated from heat-flow data yield similar information, as does thermobarometry of upper-mantle conditions obtained from kimberlite intrusions[10]. Clearly, the craton has a deep, cool and chemically reduced lithospheric root[11], comprising about 200-420×10^6 km^3 of tectosphere, a large proportion of which is Archaean in age[11].

At surface, the craton can be subdivided into a number of Archaean subdomains[12-14] (Fig. 1*a*). Teleseismic P- and PkP-wave travel-time residuals suggest that some of the subdomains are also well defined at depth, and that local changes in depth to the bottom of the lithosphere correspond closely with subdomain boundaries at surface (E. Peberdy, thesis in preparation). The oldest known subdomains occur in the eastern region of the craton: this is the Ancient Gneiss terrain and the southern Barberton terrain (Fig. 1*a*). The comagmatic mafic and ultramafic rocks of the southern Barberton terrain represent a remnant of very early oceanic lithosphere (~3.5 Gyr) which was obducted, roughly 45 Myr after its formation, onto a volcanic arc-like terrain by processes similar to those which have emplaced modern ophiolites at convergent margins of Phanerozoic continents[15-20]. This period of thin-skinned thrusting (3.4-3.6 Gyr) in ocean and arc-like environments was followed by amalgamation of the ocean and arc-like terrains at 3.2-3.3 Gyr; the region finally stabilized at ~3.1 Gyr (refs 15, 19). Careful consideration of geochronological data from the Barberton and surrounding region thus allows modelling of the first emergence and stabilization of a continental area in excess of 5×10^5 km^2; we refer to this earliest continental region as the Kaapvaal shield, or simply as the shield (Fig. 1, Table 1). Mantle xenoliths from kimberlites indicate that most of the lithospheric mantle of this shield consisted of metamorphosed peridotites and basalts (eclogites)[11], and surface geology has shown that the overlying crust was a mixture of mafic and ultramafic (greenstone), and tonalitic-trondhjemitic (granite) rocks, at various metamorphic grades[12]. Geochemistry of Archaean and Proterozoic sediments derived from this shield indicates that the near-surface proportion of greenstone to granite must have been roughly 3:7 (ref. 21). Surface exposure of the preserved crust indicates that this ratio is too high (see Figs 1*b* and 4*b*). Geophysical observations also support a lower ratio: seismic reflection and refraction data across the central craton, combined with the geological observations, indicate that the Archaean crust-mantle boundary transition was at ~35-40 km; and because P-wave velocities exceeding 7 km s^{-1} are only found in the lowermost 10% of this crust[22], a substantial upper crust of dominantly felsic composition is evident. A considerable fraction of greenstone material must therefore have been eroded from the upper crustal levels of the craton.

Along the northern margin of the craton, granulite terrains of very different crustal architecture have been tectonically juxtaposed with low-grade granite-greenstone terrains, during two successive late Archaean episodes[23-25] (Fig. 1*a*). First, northerly directed thrusting has emplaced the high-grade 'continental-shelf' terrain of the central Limpopo belt and a high-grade granite-greenstone terrain of the northern Limpopo belt on the low-grade granite-greenstone terrains of the Zimbabwean craton[23-26]. In the southern part of the Limpopo belt, a shield-type granite-greenstone terrain was similarly thrust northward. Single zircon U-Pb analyses indicate that this northward-directed tectonic transportation must have taken place between 2.7 and 2.9 Gyr (ref. 27). At a later stage, southerly directed thrusting, as well as east-west transcurrent shearing, occurred during the Limpopo orogeny (2.675±0.05 Gyr). These two periods of compressive deformation can be reconciled, respectively, with the development of a North American Cordillera-type accretionary orogeny of composite terranes along the northern margin of the shield[14,28], and with later overprinting by an alpine or himalayan-type continent-continent collision[23-25] (Table 1).

Between the period of the crustal amalgamation (3.2 Gyr) of the shield and the end of the Limpopo orogeny, several important tectonic events affected the central part of the shield. These tectonic processes are reflected in the depositional environments of major overlying Archaean sedimentary basins (Fig. 1).

Mid-Archaean shield development (3.7–3.0 Gyr)

The core region of the Kaapvaal shield consists of at least three well defined tectono-stratigraphic terrains: the Ancient Gneiss terrain, the southern Barberton greenstone terrain and the northern Barberton greenstone terrain (Fig. 1). The boundary zones between these terrains are all tectonic, characterized by thrust and strike-slip faults[14,15,19,29-31]. The two Barberton greenstone terrains, which together form the Barberton greenstone belt, are part of a fold-and-thrust belt with rocks ranging in age between 3.54 and 3.08 Gyr (refs 14, 15, 19, 20). The Ancient Gneiss terrain to the south contains some older tonalitic rocks (up to 3.65 Gyr)[32-34].

Mafic and ultramafic rock sequences of the southern Barberton greenstone terrain, dated at ~3.5 Gyr, are similar to parts of Phanerozoic ophiolites[16]. But the degree of hydrothermal overprint is considerably greater in this inferred Barberton ophiolite (referred to as the Jamestown ophiolite complex, JOC; ref. 16), and is associated with considerable SiO_2/MgO metasomatism and black-smoker-like mineralization. The hydrothermal overprint occurred during processes analogous to those occurring at mid-ocean ridges, but at greater heat and fluid flux than today[15,16,30,35,36]. The black-smoker-like mineralization was precipitated from saline brines at temperatures between 90 and 150 °C (refs 15, 37, 38). Oxygen isotope signatures indicate that the H_2O responsible for the precipitation of associated quartz is isotopically indistinguishable from that of modern sea water[16,37,38]. Fluid inclusion studies suggest shallow marine conditions, possibly no more than 60 m, during this mineralization[37,38]. Moreover, the postulated hydrothermal discharge vents associated with the mineralization can be traced into banded iron formation, which contains postulated subareal mudpool structures[38,39]. Sedimentology and geochemistry of associated rocks have also indicated shallow-water environments[40,41]. These observations could be used to infer (if we apply a modern-day analogy) that the spreading ridges in the Barberton area were very shallow to subareal.

Between 3.46 and 3.43 Gyr, within 50×10^6 years of its formation, the JOC was thrust onto an active arc-like terrain[15,19,20,42], with a trondhjemitic-tonalitic basement which dates back to 3.54 Gyr (ref. 20; Fig. 2). Large-scale imbrication of the oceanic-like rocks also occurred before and during its emplacement[15,17,18,30,42]. A second period of major thrusting (directed northwest) affected all three terrains between 3.3 and 3.2 Gyr (refs 15, 19, 20, 43, 44; Fig. 2). The time of the second thrusting event has been constrained in the central part of the southern Barberton terrain by high-level porphyries (dated at 3.229 Gyr) that are deformed by the thrusting, and similar porphyries (dated at 3.227 Gyr) that cross-cut the thrusts[15,43]. In the northwestern region of the Barberton terrain this tectonism was accompanied by a second episode of extensive, intermediate, arc-like igneous activity[15,19,20,44]. The three terrains must have amalgamated by this time because small, chemically similar trondhjemitic intrusions of this age occur in all three terrains[15-19,43,44]. After amalgamation, transcurrent shearing occurred while regionally extensive granite sheets (the Mpuluzi intrusions and Nelspruit granite) were emplaced between 3.15 and 3.07 Gyr across the three terrain boundaries[15,44,45]. Thereafter, radioactive isotopic systems of the supracrustal sequences of this part of the craton were not pervasively disturbed[39]. Resetting of the Rb/Sr, Sm/Nd and Pb/Pb systems at ~2.7 Gyr has, however, been detected in some thin but regionally extensive carbonate horizons and altered volcanics zones, indicating focused fluid activity at elevated temperatures (≥200 °C) at that time[46,47]. A distinct $^{39}Ar/^{40}Ar$ mineral plateau at ~2.7 Gyr has also been obtained from the central Barberton terrain[48].

To the south of the Barberton region, greenstone fragments of the southern (Natal) granite-greenstone terrains (Figs 1 and 2) range in age between ~3.4 and 3.2 Gyr: the oldest fragments occur in the north and the youngest along the southern margin of the craton[34,49,50]. The southernmost fragments contain dominantly mafic rocks, but include boninites and komatiites (A. Wilson, personal communication), which have been contaminated with older 'continental' crust[50]. Shallow marine conditions are inferred from studies on associated sedimentary rocks[34]. Widespread pre- to syntectonic plutonism in this region is of the same age and composition as that of the second episode of intermediate igneous activity affecting the Barberton terrains to the north (3.3–3.2 Gyr), and was similarly followed by post-tectonic granite emplacement before 3.0 Gyr (refs 12, 34, 45). The Natal granite-greenstone terrain and parts of the Ancient Gneiss terrain are unconformably overlain by the relatively underformed volcano-sedimentary sequence of the Pongola basin (Fig. 1*b*), the onset of which was before 2.95 Gyr (ref. 34). It is therefore likely that by 3.2 Gyr the Natal terrains had also amalgamated with the Barberton terrains, and that this combined terrain had stabilized before 3.0 Gyr.

Mafic and ultramafic rock sequences along the northern edge of the Kaapvaal shield, in the Pietersburg greenstone belt, resemble the oceanic rocks of the Barberton area and are probably of similar age[19,51]. The Pietersburg rocks are believed to be part of a much larger allochthonous ophiolitic terrain, which may include similar rocks of the Giyani belt (Fig. 1*b*) and possibly the high-grade greenstone remnants of the southern marginal zone of the Limpopo belt (Fig. 1)[25,51].

In summary, granite-greenstone basement of similar age and composition to those preserved in the Barberton region exists over at least 5×10^5 km^2: it extends to the south in Natal, to the north at least as far as Pietersburg and Giyani, and to the west at least as far as Vredefort (Fig. 1). This estimate is a minimum, because the craton is only a fragment of a craton that was once much larger before the breakup of Gondwana[52]. Evidence from inclusions in diamonds from Cretaceous kimberlites, which traversed the craton like superdeep 'drill-holes', indicates that by the time of amalgamation recorded in the Barberton area (3.2–3.3 Gyr), a lithospheric 'keel' to this region was already at least 150 km thick[53,54]. From the observations of this granite-greenstone terrain, we suggest that the mechanisms of lithospheric growth in the mid-Archaean era could be similar to those that currently operate along mid-ocean ridges and along oceanic convergent margins. But the interaction between these tectonic processes may require a geodynamic framework slightly different from today's, as suggested by the large volume of the granitoid component of the crust of this shield and the thickness of its lithospheric mantle. We analyse this further below, and present our model for this new framework before we continue to trace the second period of Archaean craton history.

Earliest continental lithosphere

Relative to most Phanerozoic ophiolites, the Barberton JOC sequence is thin (<3.5 km)[16], and this implies either that (locally at least) the ~3.5-Gyr oceanic-like crust was thin[16,54] or that there has been considerable, but as yet undetected, tectonic thinning.

This ancient oceanic-like crust contained a large ultramafic component (25%) which was pervasively (>95%) hydrated, with H_2O contents ranging between 2 and 15 volume % (refs

16, 35, 36, 54). The assemblage is thus calculated to have had a density as low as ~2.67 g cm^{-3} (ref. 54). Such intensely hydrated, Mg-rich oceanic crust and upper mantle, as represented by the JOC, could thus have resisted a recycling (subduction-like) process, leading instead to obduction-dominated tectonics[54]. The earliest continental nuclei may therefore have originated through regional intraoceanic obduction, giving rise to stacking, tectonic loading and subsidence of the hydrated oceanic thrust stacks[17,18,30,36,42,54] (Fig. 3*a*, upper panel). It is postulated that stacking of ocean slabs would have resulted in melting of the lower parts of the thrust to yield extensive trondhjemite-tonalite melts under variable pressure, temperature and H_2O conditions at depths of 20 km or more[42,55] (Fig. 3*a*, lower panel). Thus, the 'essential ingredient' needed to drive this obduction-dominated tectonism and granitoid formation is a buoyant and hydrated oceanic crust: an elevated MgO and H_2O content would provide this in the form of the low-density mineral serpentine, which dominates the mineralogy of the Barberton ophiolite[16].

It is generally accepted that the Archaean oceanic crust had a much higher MgO content than oceanic crust today[56,57]. Such a crust, however, would have remained relatively dense unless H_2O had efficient low-temperature access to the MgO-bearing minerals (such as olivine and pyroxene) to convert them into low-density serpentine minerals. Furthermore, H_2O must be added to the oceanic crust to facilitate the formation of granitoid magmas[58]. The most efficient process known for hydration of mantle material takes place at mid-ocean ridges (MOR) during the formation of oceanic lithosphere. We speculate, therefore, that before 4.0 Gyr the average MOR stood above sea level, so that 'dry recycling' of the Earth's oceanic lithosphere prevailed, and therefore little continental crust formed (Fig. 3*b*). Such a model is compatible with the geochemical observations that the earliest preserved mafic-ultramafic rocks on Earth were derived from mantle sources that already had a long history of relative chemical depletion and extraction of even earlier, but recycled oceanic crust[59-61]. As a corollary, in our model the formation of the earliest continental regions was related to a time that the Earth's MORs 'drowned' below sea level, triggering the first efficient and continuous interaction of our planet's hydrosphere and mantle. Although this may have been a gradual process, it nevertheless must have been an important revolutionary episode in the history of Earth's differentiation. Because of the Earth's decreasing internal heat production, with time the average oceanic crust has gradually evolved to a more tholeiitic composition, with less Mg and more Fe than that of the Archaean oceanic crust[57]. Concomitantly, densities of the oceanic crust must have increased[57]. We believe that this would have gradually heralded the return from oceanic obduction to subduction (Fig. 3*b*).

The initial rate at which the differentiation (continental crust formation) may have taken place is not known, but in our model this must have been related to prevailing rates of oceanic lithosphere formation, which is believed to have been up to 2-3 times as great as today[57]. The immediate products of these hydrosphere-mantle interactions yielded granite-greenstone nuclei such as the Kaapvaal shield, which, together with their depleted dunitic-harzburgitic keels (Fig. 3), provided continental kernels for further (plate) tectonic accretion and collisions.

If our interpretations are correct, then the occurrence of similar rocks in the wider craton framework implies that the later (3.2-3.3 Gyr) processes of amalgamation of granite-greenstone terrains, and further internal melting of the nucleus to yield the late high-level granites (crucial for transporting radioactive heat-producing and other large-ion lithophile elements from the shield's lower crust to its upper crust, as documented in detail in refs 45 and 62), all had an important part in the final consolidation of the shield by the end of mid-Archaean times. By 3.1 Gyr the shield was firmly established and constituted a framework that could support a series of late Archaean sedimentary basins (for example the Witwatersrand, Pongola and Ventersdorp basins) which have all the characteristics of a wide variety of modern sedimentary basins (see below). In the phraseology of geologists, geophysicists and geochemists, therefore, the Kaapvaal shield was a true continent (Table 1).

Late Archaean craton development (3.1–2.6 Gyr)

The quest to decipher further the continued evolution of the Kaapvaal shield into a stable cratonic region is aided, in southern Africa, by a vast and thick succession of late Archaean sediments that overlie the rocks of the shield. These sediments have preserved a rich chronological record of the processes of further continental growth. Advances in 'reading' this stratigraphic record, particularly using U-Pb zircon studies, have recently been made. Both in the centre and along the peripheries of the craton, these sediments can be observed in direct contact with their underlying basement; here, any late Archaean tectonothermal processes that have affected the shield rocks can be equated with those preserved in the late Archaean sedimentary successions. We first review these data below and then use the information to model the consolidation of the craton.

Sedimentary basins. Following the event of crustal amalgamation at ~3.2 Gyr and further lithospheric stabilization documented in the southern sectors of the craton, the shield area suffered a period of extension reflected by extensive volcanics at the base of regional sedimentary basins which covered large sectors of the shield[12,63-67] (Fig. 1*a* and *b*; Table 1).

The Dominion Group volcanics heralded the start of the Witwatersrand sedimentary basin[12,67-69]. Stratigraphic studies indicate that the Witwatersrand basin is a polyphase basin composed of at least two cycles[63,67]. The lower part of the Witwatersrand basin probably formed on a continental platform of a passive continental margin, facing open-ocean conditions to the south and the west (W. E. L. Minter, personal communication). Zircon geochronology indicates that the basin must have been initiated between 3.07 and 3.12 Gyr and supplied with sediments from a predominantly northern source ranging in age between 3.07 and 3.35 Gyr (refs 67, 68). This lower part of the Witwatersrand group is of the same age as the post-rift, thermal subsidence of the Pongola basin along the southern margin of the craton[69,70] (Fig. 1*b*). Stratigraphy and isotope work on Pongola carbonates indicate that the southernmost outcrops of this basin were also part of a passive continental margin facing open ocean conditions to the south (ref. 70 and P. E. Matthews, personal communication).

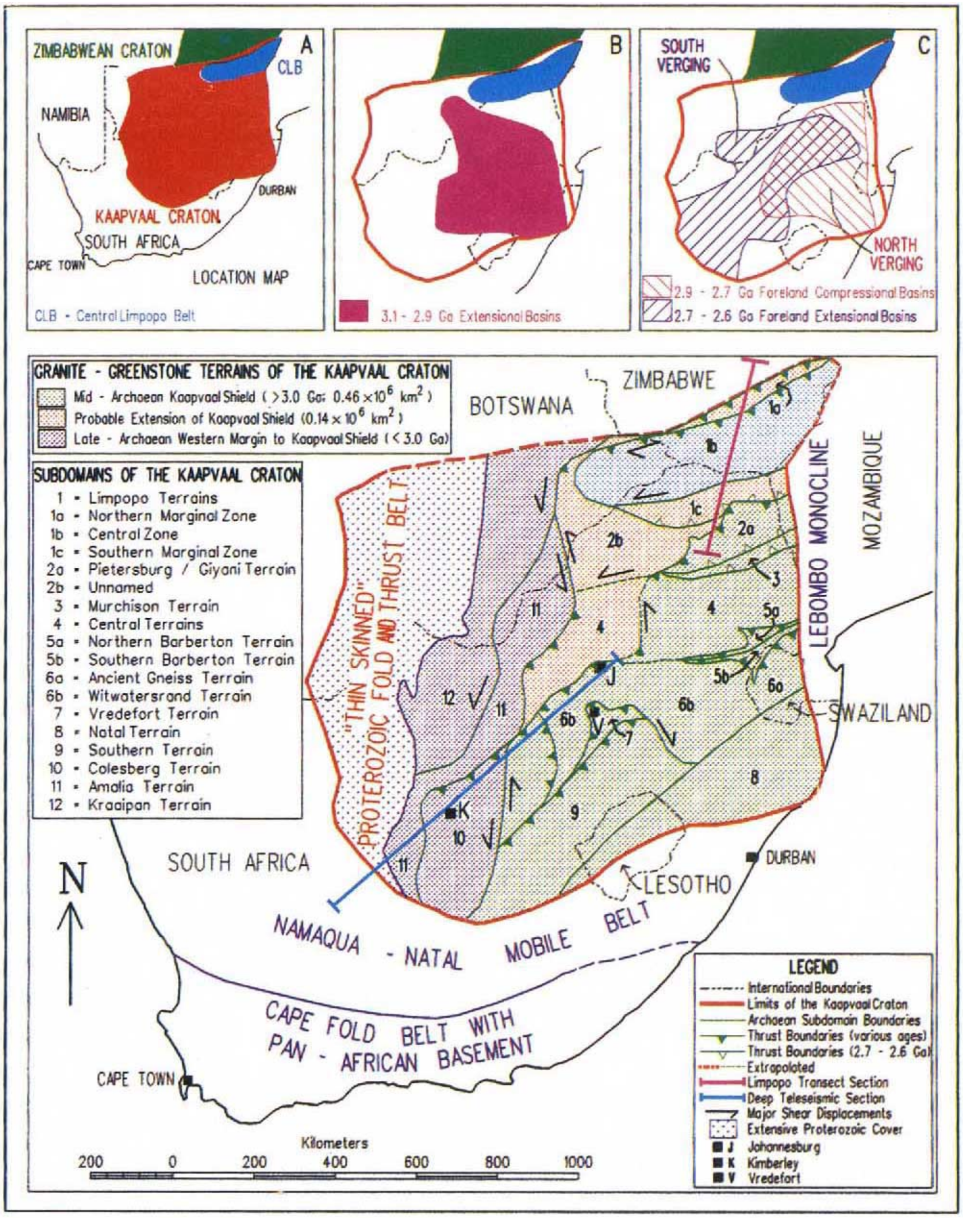
The upper part (second cycle) of the Witwatersrand basin formed in an environment influenced by convergent tectonic activity[12] and probably represents a polyphase (successor) foreland basin type[63,64], formed between 2.71 and 2.84 Gyr (ref. 67 Fig. 1*a*). This was followed by another episode of major extensional tectonics expressed by the Ventersdorp volcanism and sedimentation, now known to have formed between 2.7 and 2.6 Gyr (ref. 67; Fig. 1*a*).

Because the depositional environment of the Witwatersrand basin is in part reflected in the thermotectonic history of the underlying basement rocks, now exposed in the Johannesburg dome along the northern edge of the basin, and the Vredefort structure in the centre of the basin (Fig. 1), we first summarize the relevant geological events in these areas.

Vredefort structure. The Vredefort structure, which is thought to be the oldest and the largest known impact crater on Earth[71,72] (albeit tectonically modified)[73], is an ideal locality in which to study the basement of the Witwatersrand basin. The structure consists of an uplifted core of Archaean crystalline rocks surrounded by a collar of Witwatersrand strata (Fig. 4). Since the postulated impact at 2.0 Gyr the structure has been tectonically tilted[73], and in the exposed northwestern parts of the structure, the Archaean crystalline rocks and the Precambrian strata are now nearly vertical in attitude (Fig. 4*a*). A northwest-southwest traverse across the structure indicates that a ~36-km section, across the entire Archaean crust and the overlying strata (Fig. 4*b*), may be exposed on surface[62].

FIG. 1 *a*, Schematic diagram (based on geological, aeromagnetic and gravity data) showing the Archaean crustal subdomains of the Kaapvaal craton referred to in the text. At present, the geological data do not let us distinguish whether these subdomains are allochthonous with respect to each other (and thus significantly displaced from their original sites of formation), or were once part of a larger coherent geological province that has subsequently been tectonically disrupted. It is therefore unclear whether the domains should be classed as terranes (allochthonous; see ref. 5). To emphasize this lack of knowledge, we refer to the subdomains as terrains. Inset A shows the locations of the Kaapvaal craton, the central zone of the Limpopo belt, and part of the Zimbabwean craton. Inset B schematically shows the extent of the late Archaean extensional sedimentary basins (3.1–2.9 Gyr). Inset C schematically shows the extent of the subsequent late Archaean foreland basins formed during extended periods of compression across the craton, culminating in the 2.7 Gyr collision between the Zimbabwean and Kaapvaal cratons. The red line (Limpopo transect section) locates the section shown in Fig. 5. The blue line shows the deep teleseismic line across the southwestern part of the Kaapvaal craton along which P-wave experiments are being conducted (E. P., thesis in preparation). The minimum depth to the bottom of the lithosphere beneath the central craton varies from 170 to 310 km; distinct changes in depth occur directly beneath the different terrain boundaries delineated at surface (E.P., thesis in preparation). *b*, Distribution of the granite–greenstone terrains of the Kaapvaal craton from surface geology[12] and aeromagnetics[6]. Note the change in regional tectonic anisotropy as delineated by the greenstone belts, from northeast–southwest in the older central and southern part of the craton, to north–south in the younger western parts of the craton. Note also that in two regions (the southern margin of the Limpopo belt and within the Vredefort structure), the granite–greenstone terrains are at granulite-grade of metamorphism, exposing deep crustal sections of the Archaean cratonic crust (see also Figs 4 and 5). Also shown are the present-day extents (from geology and geophysics) of the Witwatersrand and Pongola basins. There is general consensus that these basins are erosional remnants of originally much larger sedimentary depositories, as shown schematically on Fig. 1, inset B (personal communications with participants of the Precambrian Sedimentary Basins conference, Pretoria, 1991).

a



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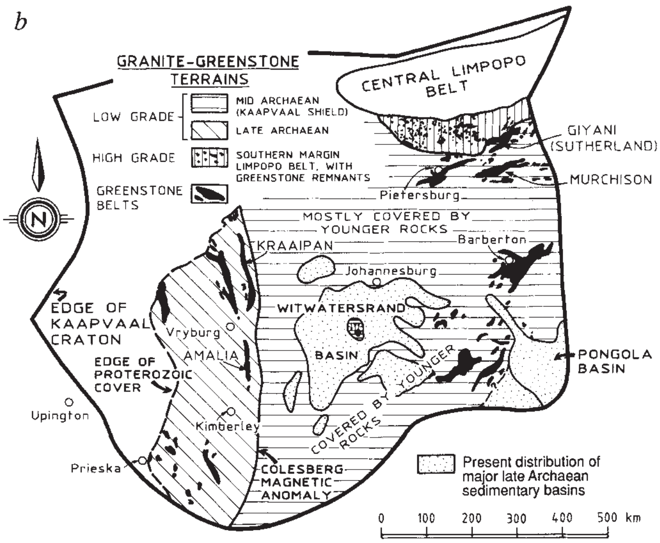


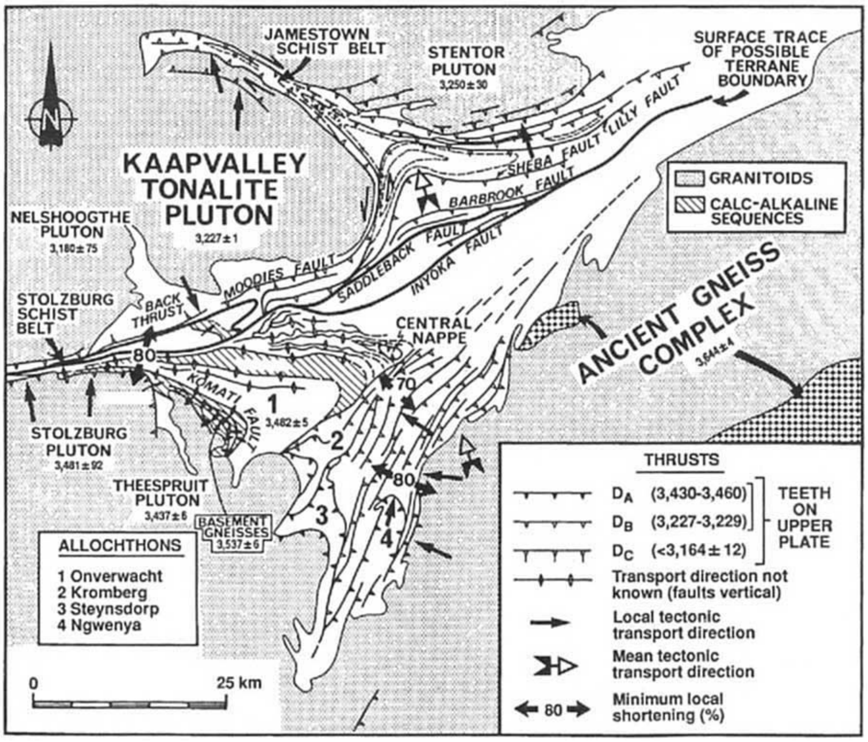


FIG. 2 Principal thrusts, strike-slip faults and allochthons now recognised in the Barberton fold and thrust belt[15,19]. A large structural break along the Inyoka-Saddleback fault system separates the different lower lithostratigraphic sequences of the northern and southern Barberton terrains of Barberton[19]. Gravity data show the belt to be ~4 km thick, with some 'roots' having a maximum thickness of 8 km (ref. 13); note that old 'continental' basement rocks occur as tectonic slices within the greenstone belt. Ages given (in Myr) refer only to thrusts shown in central zone of the belt, immediately south of the surface trace of the possible terrain boundary; many of the northeast-southwest thrust faults may in fact be of D_2 age (ref. 15).

The evidence offered by the Vredefort profile indicates that tectonic stacking of thick (>10 km) continental lithosphere slabs played an important role in the evolution of the craton. In particular, the transition from upper-crustal amphibolites to lower-crystal granulites is characterized by a brittle-ductile shear zone known as the Vredefort discontinuity[62,73]. Structural and geochemical data indicate that between 2.5 and 2.8 Gyr, the amphibolitic upper granite crust (outer gneiss granite, or OGG) was tectonically juxtaposed above granite-greenstone crust (Inlandsee leucogranofels, ILG) along an extensive thrust plane located ~8 km beneath the base of the Witwatersrand basin (Fig. 4*b*). Vibroseis data has shown that this thrust plane is of regional extent[74].

The lowermost rocks of the steeply dipping section, ~36 km into the section, consist of serpentinized harzburgites[75]. These harzburgites are thought to represent uppermost-mantle rocks of the craton; geochemical evidence suggests that these rocks were once part of Archaean oceanic lithosphere and may be equivalent to parts of the ~3.5-Gyr JOC exposed in the Barberton terrain[75]. The harzburgites were tectonically emplaced beneath the granitoid rocks and metamorphosed to ultramafic granulites, consistent with a model of Archaean tectonic, as opposed to igneous, thickening of the shield[17,18,30,54].

Johannesburg dome and northern greenstone belts. Principal north-verging thrust faults along the southern edge of the Johannesburg dome interleave granite-greenstone basement with Witwatersrand sediments[76,77]. This tectonism can be placed within the time bracket of 2.4-2.8 Gyr, that is, within the range of the thrusting recorded at the Vredefort structure. Similarly, field work and zircon dating in the Pietersburg and Giyani greenstone belts, situated along the northernmost edge of the low-grade granite-greenstone terrains of the craton (Fig. 1), has shown that major northward thrusting along the northern margin of the craton occurred between 2.7 and 2.9 Gyr (refs 19, 27, 51, 78). Coarse clastic terrestrial foreland basins, preserved in the Pietersburg and Giyani greenstone belts, formed syntectonically with this northward-thrusting, riding 'piggy-back' on the top of main thrust zones[19,51]. These basins are therefore age-equivalents of the upper Witwatersrand basin and, like the upper Witwatersrand basin, their sediments were derived predominantly from the north[19,27,51].

In summary, there seems to be a regional tectono-stratigraphic link between the shield's granite-greenstone terrains from at least as far south as the Vredefort terrain to the edge of the Limpopo terrains, through a regional north-verging foreland basin (or series of intermontane 'piggy-back' basins). How do these events relate to the orogenic activities of the Limpopo belt along the northern margin of this shield?

Limpopo belt. The Limpopo belt provides a complete section from the edge of low-grade granite-greenstone terrain of the shield to the high-grade granulite terrain and then to the low-grade granite-greenstone terrain of the Zimbabwean craton[23-26] (Fig. 1). The granulite terrain has been interpreted as the root zone of an orogenic belt formed during the collision of the two cratons in late Archaean times[12,23,26]. The belt can be subdivided into three main subterrains: the southern marginal, the central and the northern marginal zones (Figs 1 and 5). Geological and Pb-isotope data indicate that the central zone constitutes a distinctly different terrain from those to the north and south, and is separated from them by tectonic breaks[23,25,79]. Data on seismic reflection, magnetic, gravity and electrical conductivity all support models that incorporate major northward-thrusting across the entire belt (Fig. 5)[13,24,25]; the central zone has also been displaced westward[79] (see also Fig. 1*a*).

These northward-directed events were followed by crustal shortening and considerable shear-zone development in the form of south-verging thrusts and northeast-southwest striking strike-slip faults[23-25,78] (Fig. 1*a* and Fig. 5). The shear-zones were directly responsible for the rehydration of overthrust granulite-facies assemblages; away from these zones the granulites preserve classic decompression textures[23,24]. The central and southern marginal zones indicate rapid uplifts of ~20 km at 2.675±0.05 Gyr (refs 23, 24). The geological data suggest that this uplift was due to the formation of a classic tectonic 'pop-up' structure[24], and that the southward thrusting can be interpreted as a strong back-thrust[24,78]; that is, part of retrocharriage such as documented in many Phanerozoic orogenic belts[80]. Retrograde granulites of the southern marginal zone occur as isolated nappes (klippen) above prograde granite-greenstone rocks up to at least 50 km south of the thrust front[24]. Beneath the Limpopo orogenic belt, the present crust thickness is similar to thicknesses beneath Palaeozoic mountain chains (30-35 km), but it thickens northwards to more than 40 km beneath the Zimbabwean cratonic foreland[24,81]. Along the southern margin there is an apparent 5-km offset in Moho depth across the major south-verging backthrust (Fig. 5). Vibroseis reflection data across the belt show a clear fanning pattern of crustal reflectors[24] similar to the characteristic fanning crustal-reflection patterns of many Phanerozoic and some Proterozoic orogens, including the Alps, Pyrenees, Appalachians, Caledonides, Svecofennides and Karelides[82]. In each of these orogens, the fanning reflection within the crust seems to mark tectonically imbricated crust on opposite sides of a suture zone. By analogy, we interpret the offset Moho beneath the southern marginal zone of the Limpopo belt as the site of underthrusting of the continental crust of the shield within the collision zone.

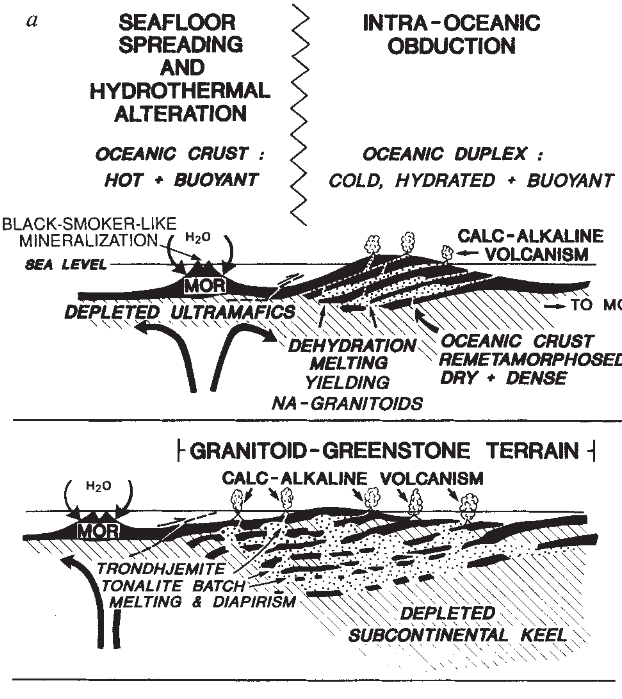
West of the Colesburg lineament. Little is known about the terrains along the western margin of the shield, to the west of the Colesburg magnetic lineament (Fig. 1*b*). Age determinations on detritus derived from these terrains as well as sporadic basement outcrops indicate late Archaean ages (2.5–3.0 Gyr)[68,69]. It has been suggested that these terrains represent the roots of calc-alkaline arcs[64]. A north–south tectonic 'grain' in this western margin is clear from the geology[83] as well as from aeromagnetics[6,84] (Fig. 1*b*). When linked with Limpopo-aged north–south shortening over the whole craton, left-lateral wrench movements must have operated along the western margin of the craton: the terrains along this margin may therefore be equivalents of the younger (2.6–2.7 Gyr) north–south trending arc and back-arc-basin terrains of western Zimbabwe[85], and could be (displaced) fragments of an extensive late Archaean, north-trending convergent boundary along the western margin of the newly amalgamated southern African shield (Kaapvaal–Zimbabwean cratons combined). The tectonics along this western margin also affected sedimentation along the western margin of the uppermost Witwatersrand basin[64].

Consolidation of an Archaean continent

The inferred time of juxtaposition of the lithotectonic terrains at lower crustal levels documented at Vredefort, and the upper-crustal northward thrusting in the Witwatersrand basin is within the age range of the extensive northward thrusting that occurred between ~2.7 and 2.9 Gyr along the northern margin of the craton described above. The available data suggest that during this period, much of this northern margin may have been assembled through a major amalgamation of elongate terrains (some of which may have been 'exotic'): examples are the Murchison belt (which is an arc-like terrain[86]), the central terrain of the Limpopo mobile belt (which is a continental fragment with associated shelf sediments[87]), and the Pietersburg, Giyani and southern Limpopo Marginal terrains (as indicated previously, these are probably composites of the early granite-greenstone terrains that constitute the shield; Fig. 1)[19,24,51].

Assembly and amalgamation of the terrains along the northern margin in the central region, and along the western margin of the craton, all overlapped with the period of sedimentation in the upper Witwatersrand basin and in the northern greenstone belts (bracketed by the ~2.7 and 2.9 Gyr deformation events

a



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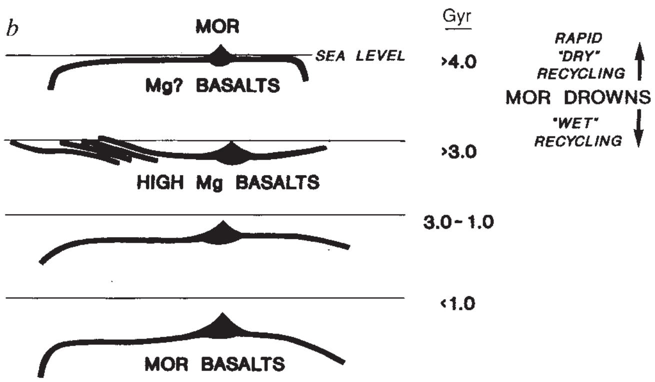


FIG. 3 *a*, Schematic model depicting the formation of the Earth's earliest preserved continental lithosphere, through linked processes of spreading and obduction-dominated tectonics. Heat flux and hydrothermal interaction of Archaean MORs was more intense than today[16,30,35,36] causing the formation of chemically enriched MgO crust (because of hotter mantle melts[56,57]), but physically less-dense oceanic lithosphere (because of associated greater degrees of hydration, and silica, iron and magnesium metasomatism[16,30,35,36]), than currently. Consequently, continuous imbrication of oceanic lithosphere led to the formation of intraoceanic thrust-stacks (upper panel). This resulted in regional load-induced subsidence (lower panel), which in turn caused *in situ* differentiation and density inversions during dehydration melting of the 'wet' oceanic-like rocks to form the vast amounts of tonalite-trondhjemites and associated calc-alkaline extrusives now present in and around greenstone belts such as those preserved in the Barberton terrains. These intraoceanic processes produced a shield area consisting of a mafic–ultramafic keel and an overlying continental-like crust by tectonic rather than igneous thickening. *b*, Schematic model for the evolution of oceanic plates with respect to sea level, and related formation of the earliest continental crust (granite–greenstone terrains) and lithospheric keel (peridotite-eclogite) of the Kaapvaal craton, following gradual 'drowning' of MORs at ~4.0–3.6 Gyr. Submergence of MORs was probably related to both the decreasing heat content of the Earth and the ageing of average oceanic lithosphere. It therefore reflects a thermally controlled increase in the average depth of the ocean basins[57,95]. No interaction between the hydrosphere and mantle occurred before this, so that 'dry' recycling of oceanic lithosphere prevailed; consequently continent formation was inefficient or absent[58]. Gradual decline in the magnesium content of the oceanic crust (caused by declining average mantle temperatures[57], and thus the decrease of the volume of low-density serpentinite, heralds the end of intraoceanic obduction-dominated tectonics and the start of modern-day subduction processes of tholeiitic oceanic crust. Although the detailed time framework for this transition is not known, Phanerozoic-like ophiolites of ~2.0 Gyr in age[82] have now been recorded, suggesting that by the early Proterozoic, oceanic processes were similar to those of today, but with a greater potential for shallow-angle subduction[95], as schematically shown in the third frame of *b*. Studies on Proterozoic ophiolites and associated sediments[82,100–102] (and P.E. Matthews, personal communication), however, indicate that modern-type MOR-basalts chemistry, and ophiolite thicknesses comparable to those of the Phanerozoic, as well as probable MOR-depths similar to the present day, were attained between 1.0 and 2.0 Gyr.

and the deposition of the Ventersdorp lavas dated at 2.714 Gyr)[67]. One of the chief problems is the observation that the northward thrusting recorded in the Witwatersrand basin (and its underlying basement) coincides with a consistently northerly and westerly derivation of the sediments[88] of the basin before the Limpopo orogeny. Clearly, the source regions for most of the sedimentation of the Witwatersrand basin cannot be directly related to the Limpopo 'highlands', and the basin is therefore not a simple foreland basin to the Limpopo belt as previously suggested[63,89]. One way to account for these observations is to model the events in two stages as follows.

Primary orogenic collage. Early (2.7–2.9 Gyr) northwards thrusting during an extended period of tectonic accretion of oceanic and continental fragments formed a wide orogenic belt along the northern margin of the shield[14,27,28]. This could be compared with the Cenozoic accretionary orogeny bounding the Canadian shield in British Columbia and southern Alaska[5,90]. In this model, subduction was to the south, and the accretion process and the subduction zone periodically shifted northwards. During this tectonism, sediments (containing detrital gold and diamonds) were shed southward into orogenic (piggy-back) basins which were overprinted by out-of-sequence northward thrusting. In such a scenario the upper Witwatersrand basin(s) and the northern greenstone-belt basins might be comparable with the (gold-bearing) intermontane basins within, and north of, the accretionary orogeny of southern Alaska.

Collision orogeny and associated extrusion, extension and collapse. Late (2.68 Gyr) southward thrusting documents the final collision of a much larger continental fragment (the Zimbabwean craton) causing backthrusting (retrocharriage) along the southern edge of the Limpopo belt and the formation of a classic 'pop-up' structure of the central Limpopo orogeny (Fig. 5), which can

a

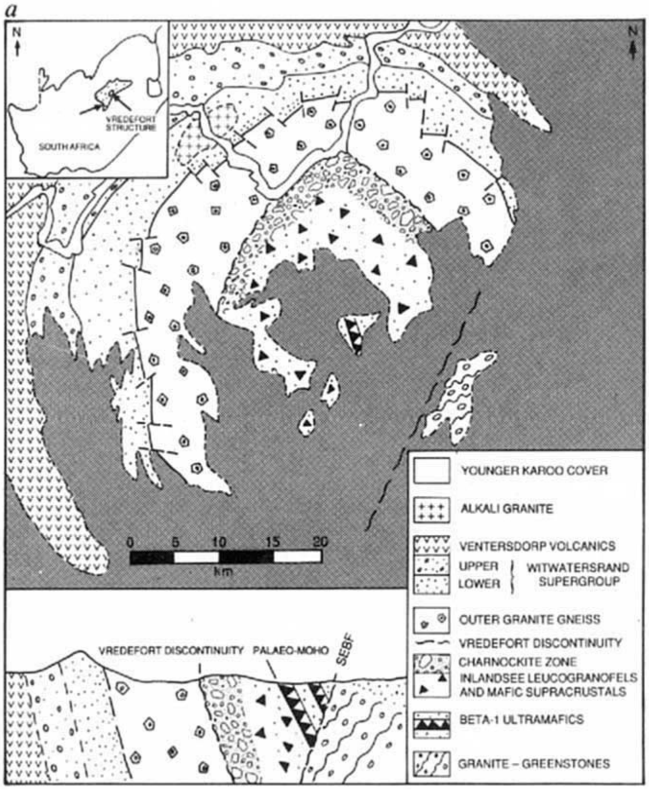
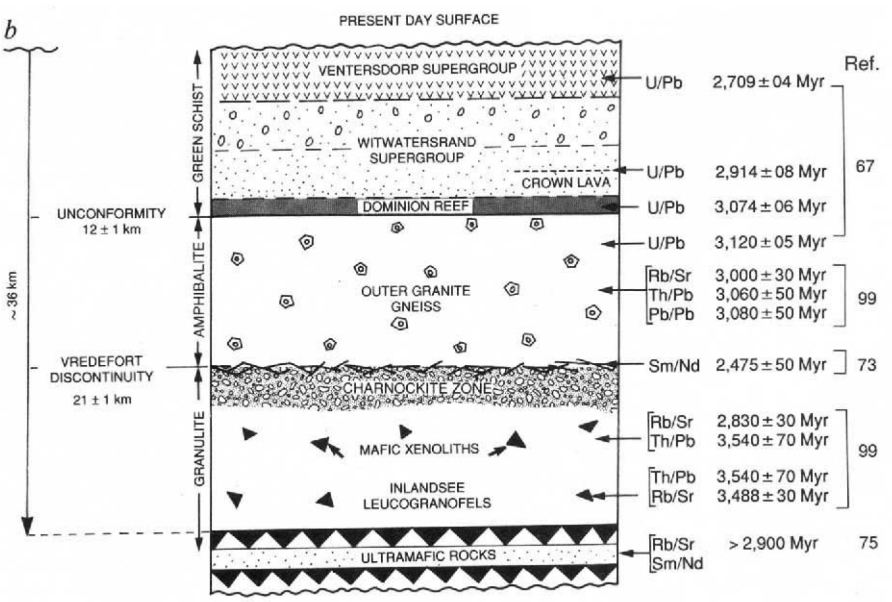


FIG. 4 *a*, Geological map and a schematic cross-section of the Vredefort structure (modified from refs 62, 73, 75; SEBF, southeast boundary fault). At least four geological domains have been identified in the basement core of the Vredefort structure. Their age relationships and stratigraphic positions in the crust, before uplift 2.0 Gyr ago, are shown in *b*. Following the section from the northwest to the southeast, the upper 10 km (in contact with the overlying Witwatersrand strata) consist of a sequence of granitic and granodioritic rocks in amphibolite facies, termed the outer granite gneiss (OGG) terrain. Beneath the OGG, a heterogenous granulite facies domain consisting of charnockites-enderbites, granulites (mafic and felsic) and supracrustal rocks is exposed. Collectively this lower crustal sequence is known as the Inlandsee leucogranofels (ILG) terrain. The OGG and the ILG are separated by a brittle-ductile shear zone known as the Vredefort discontinuity. The deepest rocks exposed near the core of the Vredefort structure consist of Archaean harzburgites which are thought to represent the upper mantle of the Kaapvaal craton. To the southeast of the ultramafic rocks, the structure is transected by the northeast-southwest-trending SEBF. Field evidence indicates that the age of SEBF is earlier than Witwatersrand. The basement to the southeast of the SEBF is a granite-greenstone domain at greenschist facies. The three different terrains (the OGG, the ILG and the low-grade granite-greenstone terrain) were probably tectonically juxtaposed between 2.5 and 2.8 Gyr. *b*, Composite stratigraphic section, with representative isotope signatures, through the crust into the uppermost Archaean mantle of the Kaapvaal craton, as represented by the Vredefort 'crust on edge' profile.

b



adequately account for both the regional decompressional metamorphic reactions of the Limpopo orogeny[23-25] and the regionally inverted metamorphic gradients along the northern margin of the craton[78]. This south-directed overthrusting juxtaposed granulites of the central Limpopo orogeny across the granite-greenstone terrain to the south; as mentioned before, granulite klippen are at present exposed up to 50 km south of the surface trace of the main south-verging thrust[24] (Fig. 5). This backthrusting was important across the whole craton: regional southward thrusting has been recorded as far south as the Murchison terrain[86], and the shield may have been affected as far south as Barberton, where rocks were disturbed at 2.7 Gyr, as recorded by the low-temperature fluid activity and resetting of the Ar–Ar, Rb–Sr, Pb–Pb and Sm–Nd isotope systems. This isotope resetting may well be related to regional tectonic-induced fluid transportation away from the Limpopo orogenic front, as has been shown for regional fluid migration through extensive sections of the crust ahead of Phanerozoic collision zones during final consolidation of major continents such as the Pangea[91,92]. Considerable crustal extension and associated crustal remelting in the form of late granite–syenite intrusions occurred throughout the craton during this period[12,44,45], as did the eruptive sequences within the extensive Ventersdorp supergroup[12,63-65,67] (Fig. 1*c*). This probably represented a period of orogenic extensional collapse[80,93]. Similarly, in this model, the westward displacement of the central zone of the Limpopo belt is probably a manifestation of extrusion tectonics (compare with refs 80 and 93) between the two colliding cratons.

The above processes mark the end of an extended period of Archaean consolidation of a large section of the shield into a true craton (that is, a stable continent).

Continental lithosphere growth

The Kaapvaal craton was established over a period of at least 1 Gyr. The first $\sim 5\times10^8$ years (3.7–3.2 Gyr) saw the emergence of a 'continental' shield area through large-scale intra-oceanic thrusting, tectonic burial and melting of hydrated oceanic crust. The establishment of this continental platform and its underlying tectosphere thus proceeded by tectono-igneous differentiation, driven essentially by the addition of water to mafic–ultramafic crust. Considering the large ratio of granitoid to greenstone in the shield[21], and taking into account the knowledge that the granitoids were derived from the greenstone equivalents (hydrated basalts of same age and geochemistry)[42,45], it is clear that ocean-floor recycling and associated tectonic processes (subduction/obduction) must have operated.

It is possible that modern counterparts of shield are to be found amongst the present-day oceanic plateaux of the Western Pacific, such as the Ontong-Java Plateau (OJP). Like the Pacific ocean plateaux, which are destined to collide with oceanic and continental arcs[5,90,94], the Kaapvaal shield apparently collided with arcs and oceanic terrains. It is therefore interesting to note that the OJP is similar in size to the Kaapvaal shield[94] and that the OJP is now colliding with, and obducting across, the Solomon Islands arc/back-arc terrain[94], which has already been specifically compared to Archaean greenstone belts[95]. The OJP has a similar crust thickness and similar seismic velocities to those postulated for the shield (35–40 km; 6–7 km s^{-1}; refs 22, 94). Kimberlite-like intrusions (alnöites) into the OJP contain deep-seated xenoliths (garnet lherzolites) indicating a thick (up to 135 km) depleted oceanic lithosphere beneath the plateau[96]. The depleted mantle beneath the OJP has been underthrust by hydrated oceanic crust[96], and the ocean plateaux in general are believed to contain high-Mg basalts and komatiites[97]. But a problem that must be considered when comparing the Kaapvaal shield to present-day ocean plateaux is that, according to our model, plateaux-like OJP should also contain large volumes of tonalites.

In the case of the Kaapvaal shield, intra-crustal melting yielded an uppermost granite 'layer' during the period of amalgamation, converting the shield into a truly continental platform of at least 5×10^5 km^2, with 40-km-thick crust and a lithosphere boundary between 170 and 350 km in depth by 3.1 Gyr. The lithospheric growth rate of the shield, given its formation between 3.2 and 3.6 Gyr, must therefore be $\sim$0.5 km^3 yr^{-1}; and the crustal growth rate must be $\sim$0.05 km^3 yr^{-1} (as opposed to the global Phanerozic crustal growth rate of 1.06 km^3 yr^{-1} calculated by Reymer and Schubert[98]). If we assume that the total crustal growth rates in the Archaean were 3–4 times the present Phanerozoic rate[1,57,95,98], then between 3.2 and 3.6 Gyr, the world's continental surface area must have increased by 64–85 times the size of the Kaapvaal shield ($\sim$15–20% of the present continental surface area). As there are at most 10 preserved Archaean cratons worldwide (of which only two substantial

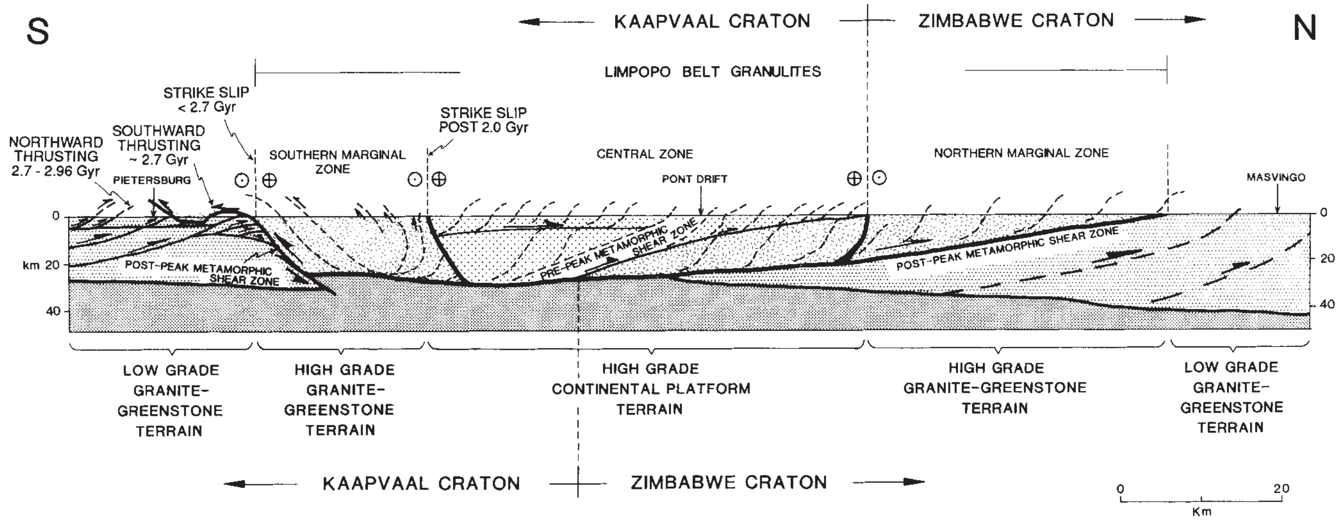


FIG. 5 A composite north–south profile across the Limpopo belt (see red line on Fig. 1*a* for line of section) based on geological and geoelectric, magnetic, seismic (refraction and vibroseis reflection) and gravity data (compiled from refs 24–26, 78). This profile is similar to those across several Phanerozoic and Proterozoic mobile belts[82]. The profile clearly shows the early northward-directed tectonic transport (dated 2.7–2.9 Gyr) overprinted by the $\sim$2.7 Gyr southward-directed thrusting related to the climax of the Limpopo orgeny. The 2.7 Gyr tectonic loading of the Kaapvaal craton by the Limpopo terrain probably induced focused fluid migration throughout the cratonic crust, at least as far south as the Barberton terrains. Dark shading represents upper mantle. ⊙, transcurrent movement out of the page; ⊕, transcurrent movement into the page.

ones are older than 3.0 Gyr), it is difficult to avoid the conclusion that Archaean continental crust was recycled back into the mantle. The implications of this will be discussed elsewhere (M.J.d.W. and W. Wilsher, manuscript in preparation).

During the second 5×10^8-yr period (3.1–2.6 Gyr), the shield extended by a further $\sim5\times10^5$ km^2. It is, however, not clear at this stage how much of this continental growth was due to a net addition of material newly derived from the mantle.

Finally, the geological manifestations of the tectonic and thermal processes that converted the shield into the craton seem to be identical to those occurring during the Phanerozoic. The kinematic details of the tectonic processes remain less well understood, because the time spans within which Archaean geological processes can be resolved are at least an order of magnitude greater than for Phanerozoic orogenic processes; and considering the complexities of the latter, this leaves a multitude of tectonic models that still need testing before the formation and growth of this Archaean continent are understood. □

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