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On Archean granites, greenstones, cratons and tectonics: does the evidence demand a verdict?

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Abstract

I review geologic evidence for Archean plate tectonic processes within the framework of geophysical, geochemical and experimental observations. For the Late Archean (2.5–3.0 Ga), the evidence is primarily based on the rich database of the Superior craton; for the Early Archean (3.0–4.0 Ga), data from the Pilbara and Kaapvaal cratons are examined. Data from other old cratons are used as supplementary evidence. The verdict is that there is a robust consensus on the validity of using plate tectonic boundary processes to decipher the Late Archean rock record; and it also confirms that such processes dominated the early Archean.

Modern plate tectonics probably has its roots in the Hadean-Archean transition between 4.0 and 4.2 Ga. Before that time, recycling of thin, dry oceanic lithosphere may have dominated the surface tectonics of Earth. Assuming gradual mantle dehydration during formation of oceanic lithosphere, from 4.5 Ga onward, ca. 70% of mantle water would have accumulated in the hydrosphere by about 4.0 Ga. At this stage, Earth's seafloor-spreading plate boundaries, would, for the first time, have come to operate below sea level, in turn, initiating onset of extensive hydrothermal cooling and alteration of upper oceanic lithosphere as observed today. Only then did 'modern' production of granitoids become possible during the recycling of Earth's lithosphere. Processes at convergent plate boundaries, however, evolved more slowly into a modern subduction-dehydration style over the course of the Archean. In the earliest Archean, recycling of buoyant hydrated oceanic materials into the mantle may not have been efficient, and back-arc spreading processes may have been absent. Instead, I speculate that hydrated oceanic lithosphere 'piled-up' to form intraoceanic thrust-stacks, which evolved into Earth's first continental fragments through internal differentiation during tectonic thickening and construction of deep lithospheric keels. These fragments amalgamated into larger cratons throughout Archean times.

Minimum continental growth rates of various cratons spanning the Archean period illustrate that Early Archean continental crustal-growth must have been 'explosive', and that only a small fraction of this crust has been preserved. I speculate that modern subduction-style recycling of exosphere materials (here taken as crust, sediments, hydrosphere, atmosphere) may not have been efficient during the Early Archean, and that many Early Archean continental fragments, possibly those lacking a thick buoyant mantle keel, were episodically 'flushed' whole-scale, back into the convecting mantle. Those cratons that survived, built up and/or retained thick Archean mantle keels (between 200 and 400 km). Delamination of these keels cannot, therefore, be invoked to explain subsequent Archean (and later) tectonic events recorded at the surface of these cratons. The style of modern subduction-related recycling of exosphere materials emerged during the Late Archean. These conjectures have major consequences for global mass balance and chemical flux considerations.

New experimental work indicates that komatiitic magmas were hydrous, and that their liquidus temperatures are comparable to those of magmas observed at modern plate boundaries. This, and other independent observations

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indicate that the upper mantle in the Archean was not excessively hotter than today. The presence of significantly dissolved water in komatiitic magmas raises the possibility that Earth lost heat more effectively through dehydration from a wetter, cooler and less viscous mantle: 'wet' Archean mantle dynamics may have been significantly different than the present 'dry' one. Since hot spot volcanism is less efficient at dissapating internal heat as through spreading at mid-ocean ridges, it is unlikely that Archean plates were swamped by plumes, or replaced by plume tectonics. Rather, accretion and hydrothermal cooling processes at seafloor-spreading plate boundaries were possibly more efficient; and deep mantle plume materials may have been deflected via the asthenosphere into these boundaries more effectively than observed today.

Inferred secular changes in Earth's tectonic processes are mostly based on qualitative statements concerning the relative increase/decrease of preserved rock types/rock associations through time. The generally accepted decline of komatiite abundance (and their average MgO contents) from the Early Archean is a good example; and so is the apparent increase in the abundance of kimberlites with time. Analysis of the rock make-up of 40 greenstone belts worldwide, for which there is reasonable quantitative data, indicates that there is no unequivocal decline in the volume of preserved komatiites, or in the maximum MgO content of ultramafic magmas from the Early Archean to the Phanerozoic. There is, however, a peak in preservation of komatiites in globally distributed greenstone belts of the Late Archean. Similarly there may have been abundant kimberlites in the Early Archean.

The accelerating pace of accurate geochronology can now sustain more reliable palaeomagnetic studies to resolve Archean tectonic displacements; and also geochemical flux studies to track Archean reservoir systems. Together with new mapping and experimental work, particularly on mafic-ultramafic sequences, these will shed new light on Archean geodynamics. But this will require interdisciplinary Lithoprobe-type studies with international participation, and the final demise of Precambrian apartheid.

1. Introduction

Granite–greenstone assemblages of Archean age have been documented on all continents; these rock suites, often subdivided into Late Archean (2.5–3.0 Ga) and Early Archean (3.0–4.0 Ga), represent remnants of Earth's earliest preserved continents. Well preserved Late Archean continental crust has a known exposed areal extent of approximately 11×10^6 km^2, whereas well preserved rocks of Early Archean age have an areal extent of less than 0.8×10^6 km^2 (about 7% and 0.5%, respectively, of the present day exposed continental surface area). No continental rocks older than 4.1 Ga have yet been discovered. Thus, there is relatively little rock record from which to directly reconstruct the workings of the early Earth, and none from the first 500×10^6 yr of its evolution. Compared to the Phanerozoic, we are literally grappling in the dark and it is no wonder, therefore, that some of the most polarized debates about Earth evolution are rooted in shaky conjectures about its first 2000×10^6 yr. There is lack of agreement, for example, about the origin and growth rates of the first continents; the structure of the oldest oceanic lithosphere; the heat content, viscosity, layering, and chemical composition of the early mantle; and the flux of primordial materials surfacing from, and/or (re)-cycling into the mantle during these early times. Yet these are key constraints from which to chart the history of earth processes.

One fruitful way to address these problems is to carefully document and compare the tectonic histories of the best-preserved Late Archean rock sequences, and to proceed cautiously back in time from there. Lately, major advances in understanding tectonic processes of the Late Archean Earth have emerged from such studies, as was clearly evident during the recent Precambrian Conference in Montreal, Canada (Precambrian'95, 1995).

2. Late Archean tectonism: evidence from the Superior Province, Canada

There is now a robust consensus on the validity of using plate tectonic principles to decipher the Late Archean rock record. The most important contributions towards this convergence of opinions are based on the results of studies focused in the first instance on the vast Superior Province of Canada (1 572 000 km^2). Significant advances have come from two unrelated approaches following bold planning amongst the Canadian geoscience community in the late 1970s and 1980s. These initiatives are yielding a steady flow of high-quality geochronological and geophysical data which is fine-tuning some firm conclusions on the origin and evolution of this, Earth's largest well-preserved Late Archean craton.

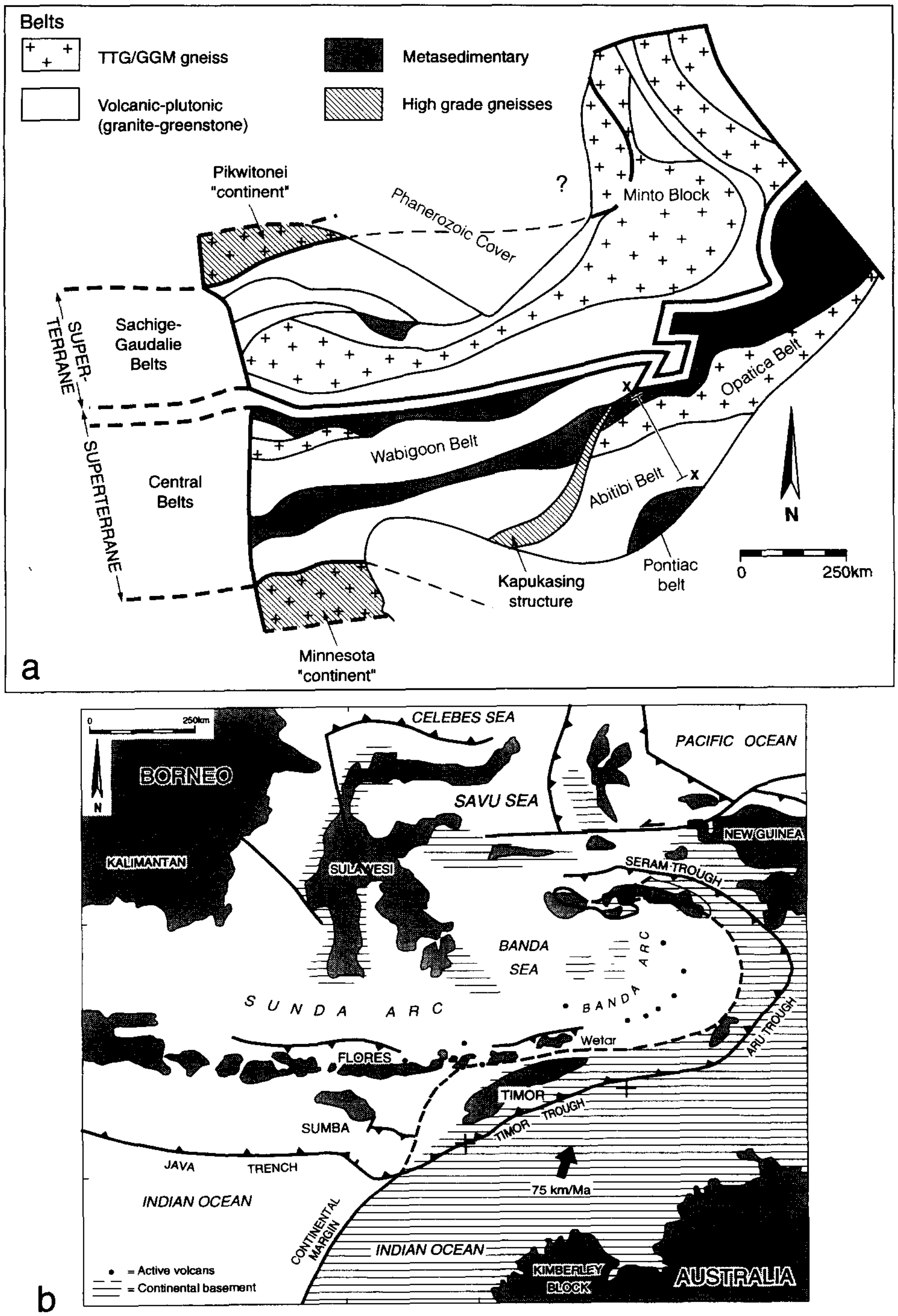
In terms of dating, the Superior Province is the best studied piece of Archean real estate in the world (Table 1); and the only Late Archean craton about which credible geochronological statements concerning crustal architecture can be made in three-dimensions. The precise single mineral U/Pb data, produced initially at the Royal Ontario Museum under the leadership of T. Krogh from the late 1970s onward, must surely be recognized as having provided the Earth Sciences with a quantum-step forward in the unravelling of early Earth processes. Some 1150 precise U/Pb single zircon dates are now published for the region (Percival J. and Davis D., pers. comms., 1996; Stott, 1997). From this data it can be deduced that the province as a whole has a history spanning some 400×10^6 yr, but that the majority (>90%) of its rocks define elongated belts (with aspect ratios of 1:10) formed between 2.7 and 2.8 Ga, and which were subsequently tectonically juxtaposed within an even shorter time span of some 30×10^6 yr, between 2720 and 2690 Ma. Canadian geologists have been advocating for some time now that this geochronology data, when integrated with their mapping and petrogenetic analysis, can best be explained by progressive north to south (present coordinates) convergent plate-boundary tectonics, culminating in crustal stacking during final convergence of two continental masses (the northern Pikwitonei-Minto and the southern Minnesota blocks; Fig. 1). A modern environment which comes to mind is that of the still evolving

Table 1
Summary of data from selected Archean cratons used to calculate their minimum growth rates

Craton	Area ($\times 10^6$ km^2)	Crustal thickness (in km) range (average)	Archean age (%) (peak (range) in Ga)		Precise U/Pb zircon ages		Net min. growth rates (km^3 y^{-1})	
			Early	Late	No. published	Density/ 10^3 km	Crust	Lithosphere
Kaapvaal (1)	1.2	32-40 (37)	60(3.6-3.2)	40(2.6-3.0)	100	0.08	0.07	0.7
Pilbara (2)	0.06	28-33 (32)	90(3.6-3.1)	10(2.8-3.0)	50	0.83	0.003	0.005
Superior (3)	1.57	35-40 (38)	10(3.0-3.3)	90(2.7-2.8)	1150	0.73	0.54	3.6
Yilgarn (4)	1.0	38-40 (39)	10(3.7-3.3)	70(2.6-2.8) 20(2.8-3.0)	150	0.15	0.14	0.29
Sao Francisco (5)	0.82	37-42 (39)	30(3.0-3.5)	50(2.7-2.9) 20(1.9-2.1)	30	0.04	0.08	0.41
Zimbabwe (6)	0.27	30-40 (35)	10(3.0-3.3)	90(2.6-2.7)	22	0.08	0.09	0.7

Data taken from de Wit and Ashwal (1997), and Refs. given below. Sources for the number of precise U/Pb zircon ages are 1996 pers. comms. (1) R. Armstrong; (2) T.E. Zegers; (3) D. Davies and J. Percival; (4) D. Nelson, and Wilde et al.,; (5) F. Baars and Teixeira et al. (1996); (6) G. Davies and Wilson et al. (1995). The densities of zircons/km^2/craton are not strictly comparable because different cratons are covered by younger rocks to different extents. The rate of production of new dates is also variable, and in some regions where both conventional and SHRIMP methods are now used (Pilbara; Superior), numbers of dates is expected to increase by ca. 100/year.

Minimum growth rates calculated for each craton represent those periods during which its greatest volume of juvenile crust/lithosphere was created (i.e. period of Early Archean for Kaapvaal, and period of Late Archean for the Superior). There are major differences in growth rates amongst different cratons both in the Early and Late Archean. This may reflect variable rates of growth or recycling of different cratonic areas. Assuming no crustal recycling, and using the highest growth rates calculated for each of the Archean periods (0.07 and 0.54 for Early and Late Archean, respectively), integrated over each period (1.0 and 0.5 Ga, respectively), we should expect at least 12 cratons of the size of the Early Archean Kaapvaal, and 172 cratons of the size of the Late Archean Superior to be preserved on Earth. This is clearly not the case, implying either extreme punctuated crustal growth (not apparent from this table), or large volume crustal recycling (see text for further discussion).


Belts
TTG/GGM gneiss
Metasedimentary
Volcanic-plutonic (granite-greenstone)
High grade gneisses
Pikwitonei "continent"
Phanerozoic Cover
?
Minto Block
SUPER-TERRANE
Sachige-Gaudalie Belts
SUPERTERRANE
Central Belts
Wabigoon Belt
Opatica Belt
Abitibi Belt
X
X
N
0 250km
Pontiac belt
Kapukasing structure
Minnesota "continent"
a
CELEBES SEA
PACIFIC OCEAN
0 250km
N
BORNEO
KALIMANTAN
SAVU SEA
NEW GUINEA
SULAWESI
SERAM TROUGH
BANDA SEA
BANDA ARC
SUNDA ARC
Wetar
FLORES
TIMOR
TIMOR TROUGH
ARU TROUGH
SUMBA
JAVA TRENCH
75 km/Ma
INDIAN OCEAN
CONTINENTAL MARGIN
INDIAN OCEAN
= Active volcans
= Continental basement
KIMBERLEY BLOCK
AUSTRALIA
b

Indonesian Archipelago as it continues to be consumed between the two continental masses of Australia and the Asian mainland (Fig. 1).

Complex facies changes abound throughout the Superior Province and remnants of juxtaposed accreted island arcs, oceanic plateaux, marginal-basin crust, forearc-prisms, and juvenile microcontinental blocks have all been identified (Stott, 1997); and most recently, melanges found in the Minto Block have provided the first documented field evidence suggestive of a nearby cryptic Archean suture-zone (Skulski and Percival, 1996).

One of the most elegant pieces of U/Pb work yet produced to corroborate Archean tectonic crustal stacking during the accretion processes, is that of T. Krogh himself (Krogh, 1993). Across a section through the Kapuskasing uplift, where geologists have speculated that one can traverse from the uppermost to lowermost crust (see Can. J. Earth Sci., 1994), Krogh has demonstrated that clastic supracrustal rocks (including conglomerates) metamorphosed to granulite grade at the lowermost levels of this crustal-section, are younger than the overlying, lower-grade volcano-plutonic rocks. Only Late Archean tectonism involving progressively deeper under-thrusting of the younger rocks beneath the older rocks, can account for these observations.

Seismic-reflection profiles obtained during the Lithoprobe programme (Clowes et al., 1993; Ludden et al., 1993) across selected transects of the Superior province, have vindicated the geological interpretations. The style of seismic reflections throughout the Archean crust beneath at least parts of the Superior Province is similar to many sections through Phanerozoic crust (Calvert et al., 1995), and its thickness (ranging between 35 and 40 km) is similar to that of Phanerozoic Andean arcs (Rudnick and Fountain, 1995).

The Superior Province is the only Archean terrain within which a fossil suture zone has been geophysically imaged. The seismic data across the transition from the Opinaca sediment belt through the arc-related Opatica belt and into the Abitibi greenstone belt, has imaged a relatively shallow, northward-dipping reflector that extends some 30 km into the mantle. This reflector is interpreted to represent a fossil 2.7 Ga suture zone along which the low grade Abitibi belt was subducted beneath the high grade Opatica belt between 2693 and 2686 Ma (Calvert et al., 1995). This, and other seismic reflection data have confirmed complex and multiple thin-skin crustal stacking beneath several of the tectonic domains of the Superior Province complementing previous crustal sections constructed using surface geology (Ludden et al., 1993).

The case seems sealed for accretion tectonics along a north-dipping master subduction zone flanking the margin of a substantial ocean. As a first approximation, the bulk of the Superior Province represents continental growth by the formation and sweeping together of a series of arcs, oceanic terranes and accretionary prisms along

Fig. 1. (a) Tectonic domain map of the main lithotectonic subdivisions of the Late Archean Superior Province, Canada (simplified after Stott, 1997). Note the presence of late Early Archean crust in the Pikwitonai and Minnesota domains; smaller fragments of relative 'juvenile' basement may also be present within some of the other domains (Corfu and Stott, 1996). The majority of the rocks within the elongate belts, however, were formed within a relatively short time span between 2.7 and 2.8 Ga and were thereafter rapidly tectonically juxtaposed within a period of 30 million years. Section X–X shows the approximate location of the seismic section of Calvert et al. (1995). This section has imaged a northward dipping band of reflections beneath the Opatica belt which extends 30 km into the mantle, and which is interpreted as a relict 2.69 Ga-old suture associated with the amalgamation of the belts (see text for further explanation). (b) Simplified map of the east Indonesian region (after Vroon et al., 1996), for comparison with the Superior Province shown at the same scale. Small 'juvenile' continental and oceanic fragments are tectonically dispersed between the elongate fore-arc, back-arc and island-arc environments of this triple junction between converging continental (SE Asia), oceanic (Pacific) and mixed (Indo–Australian) plates. Provenance areas of the small continental fragments have been identified using isotopic signatures (Vroon et al., 1996), as seems possible too for fragments embedded within the Superior Province (Corfu and Stott, 1996). Forward modelling (Vroon et al., 1996) shows that in about 10 Million years from now the Indonesian region would from a broad 'Banda Orogen' with similar elongate belts and structures to those preserved in the Superior Province, including the possible uplift and exposure of crosscutting high grade belts like the Kapuskasing structure (see Fig. 4 of Vroon et al., 1996). It is interesting to speculate that the lengths of the plate margins and the widths of the plate boundary zones of the Superior Province precursors might have been similar to those within the Indonesian region.

evolving plate boundaries, which probably originated adjacent to an older (ca. 3 Ga) continental margin (Hoffman, 1989; Feng and Kerrich, 1992; Kimura et al., 1993; Percival et al., 1994; Stott, 1997). However, other work has already shown that this is but a very simplified version of what happened in detail (Corfu and Stott, 1996). The entire Superior Province may be viewed as a remnant of one type of Archean continent, affected by a wide range of orogenic processes, including extrusion tectonics and late extensional collapse.

Whereas integrated work on other cratons is not as advanced, results from areas consisting of predominantly Late Archean sequences, such as the Yilgarn Craton (Australia), the Zimbabwean Craton, the Tanzanian Craton, the Sao Francisco Craton (Brazil), the Slave Craton (Canada), the Dharwar Craton (southern India), the western part of the Kaapvaal Craton (South Africa), the Aldan-Stanovik Shield (Siberia), the Hebei Craton (China) and the Baltic Shield, have yielded many similar tectonic interpretations (Rollinson, 1993; Rollinson and Blenkinsop, 1995; Treloar and Blenkinsop, 1995; Myers, 1993, 1995; de Wit and Ashwal, 1995; Choukroune et al., 1996; Chadwick et al., 1996; Wilde et al., 1996; Smith et al., 1998; Dirks and Jelsma, 1998 [for detailed overviews of recent work on these cratons, see various authors in de Wit and Ashwal (1997)]. In addition, the classic Nuuk region of Southern Greenland, comprising substantial remnants of amongst Earth's earliest preserved Archean rocks, is now also recognized to have amalgamated into a large cratonic segment during Late Archean assembly (Friend et al., 1996).

The evidence from the Late Archean demands, therefore, a general verdict: plate-boundary processes similar to today operated during Late Archean times.

To obtain a better global-scale perspective of Late Archean plate boundaries, however, the time has come to critically explore broader questions in which the results of the Superior province studies can be accurately compared against those from elsewhere. Was the Superior Province once part of a more extensive orogenic belt? Does, for example, the Superior Province represent part of a greater, linear assembled continent? The Neoproterozoic accretionary complex of the Arabian-Nubian Shield, part of the extensive Mozambique mobile belt, may be as a possible analogue of an Archean Craton (Harris et al., 1993; Behre, 1997). Similarly, along the Alpine-Himalayan chain extensive accretionary terrains, such as those between Greece and Pakistan, are well-preserved. Or, perhaps the Superior Province was part of an even larger accretion complex, such as the Altaids (cf. Sengor et al., 1993; Dobretsov et al., 1995; Sengor and Natal'in, 1996; Burke, 1997). Answers can only come from detailed comparative studies between the Superior Province and other Late Archean cratons where rapidly increasing geochronological databases are now starting to emerge (Table 1). Many of these Late Archean cratons have rifted margins so they do not represent their original cratonic extents (Isachsen and Bowring, 1994). Were any of these cratonic fragments once contiguous? Can the tectonic events recognized on one craton be confidently correlated to other cratons? For example, the Zimbabwe craton has a similar geological history to that of the Superior province: small remnants of Early Archean tonalitic basement (the Tokwe block; 2.9–3.3 Ga) are overlain and intruded by widespread bimodal volcanics–plutonics, developed within a relatively short period between 2.6 and 2.7 Ga (Jelsma et al., 1996; Blenkinsop et al., 1997). However, their tectonic histories appear to be different, unless this is merely a scale deception: the Superior Province is almost an order of magnitude larger than the Zimbabwean craton (Table 1). Why are the granites of some Late Archean cratons, of similar age and inferred tectonic history, dominated by Tonalite-Trondhjemite-Granodiorite (TTG) whilst others are dominated by Granodiorite-Granite-Monzogranites (GGM), such as the Superior province and the Yilgarn craton, respectively?

To answer the above questions with appropriate resolution, bold new international programmes, incorporating palaeomagnetism and a variety of integrated isotope studies, will need to be drawn into the array of ongoing field studies (see below).

One of the reasons why the Superior Province is preserved at all is almost certainly because this terrain, like all other well preserved Archean terrains, has a thick, cool, strong and highly viscous, but relatively light (buoyant, less dense than asthenosphere), mantle keel (tectosphere of Jordan, 1975, 1979, 1988), with its lower portion within the diamond stability field. Although the origin of these keels, which vary between 200 and 400 km in thickness, is not well understood (Pollack, 1986; Anderson, 1995; Walker et al., 1995; Carlson et al., 1996; Hart et al., 1997; and see below), they occur beneath all well-preserved Archean terrains of substantial size (Durrheim and Mooney, 1994; Anderson, 1995). Did all Archean continents have such keels? Perhaps only those continental fragments which attained and/or retained such keels remained protected from later tectonic reworking and recycling, and are thus the only representatives of a once much wider variety of Archean continental lithosphere than is presently preserved? At any rate, since such keels do not appear beneath any known younger tectonic terrains, they must be telling a specific Archean tale. We will return to this in a later section. For the moment we can feel confident about extrapolating plate tectonic boundary processes back as far as 3.0 Ga. What then about the Early Archean?

3. Early Archean tectonism: evidence from the granites.

The oldest rocks discovered to date are the granitoid Acasta gneisses within a small enclave ($<20\ km^2$) near the western extremity of the predominantly Late Archean Slave Craton (Bowring et al., 1989; Bowring and Housh, 1995). The oldest samples of these rocks are just over 4.0 Ga in age, and have variable isotopic signatures with both enriched and depleted ϵNd values (Bowring and Housh, 1995; Bowring, pers. comm., 1996). Although the Acasta gneisses have been deformed and metamorphosed at elevated temperatures and pressures (upper amphibolite grade), and locally re-equilibrated at about 3.4 Ga (Moorbath et al., 1996), Bowring and his colleagues have shown that Earth's oldest preserved rocks were continental and derived from hydrated, chemically depleted mafic precursors, as well as from earlier granitoids; and that the 4.1–4.0 Ga Acasta gneisses closely resemble the kind of rocks that underlie Andean-arc volcanoes, both in composition and structure.

Similar gneisses often referred to as 'grey gneisses' (Martin, 1994) have been found in small areas ($<3000\ km^2$) elsewhere around the world and represent the remainder of the Early Archean period: the 3.7–3.9 Ga Amitsoq [W.Greenland], Uivak [Labrador] and Napier [Antarctica] gneisses; the 3.6–3.7 Ga Ancient [Swaziland] and Narryer [W. Australia] gneisses; the 3.6 Ga Minnesota [USA] gneisses; the 3.4–3.5 Ga gneisses from Wyoming [USA], the Limpopo region [southern Africa], southern India and the Ukrainian shield; the 3.0–3.2 Ga Tungurcha [Siberia] gneisses, to mention a few. Each reveals ambiguous tectonic tales; some are associated with highly deformed and metamorphosed volcanic rocks, others with metamorphosed platform-like sediments such as quartzites and carbonates. Many of these gneisses have been repeatedly metamorphosed in the Late Archean–Proterozoic at granulite facies, and it is not always clear, therefore, what kind of granulite facies, if any, prevailed during the Early Archean (e.g. Grau et al., 1996). Where Early Archean metamorphism has been geochronologically resolved, however, the metamorphic conditions have yielded valuable insights into local continental transient geothermal gradients during that time: all the temperature conditions of these high-grade metamorphic rocks, irrespective of their pressure conditions, exceed the stable continental geotherm required to maintain the diamond stability field (Percival, 1994). Distinguishing between a number of possible active continental tectonic environments in which the metamorphism must have occurred is not yet possible, although the deep roots of magmatic arcs and/or collisional belts are the most favoured analogies (Percival, 1994).

Some of the oldest of these granitic gneisses, the Amitsoq gneisses, are juxtaposed with 3.7–3.8 Ga supracrustal rocks of the Isua belt and Akilia association (these sequences are now collectively referred to as the Itsaq ('ancient thing') complex; Nutman et al., 1996). The Isua sequence contains recognisable sedimentary rocks (turbidites, marbles, quartzites, conglomerates, banded iron formation) and volcanic/volcaniclastic rocks (felsic and mafic, the latter including pillow lavas), which tell something concrete about near surface processes at that time. However, these supracrustals are of very limited areal extent (less than 75 km^2) and have been thoroughly deformed at amphibolite facies conditions during the Late Archean (Nutman et al., 1984, 1996; Nutman, 1997). Thus, while some have suggested that these rocks represent a modern subduction-related accretion complex similar to those found in Japan (Maruyama et al., 1991), the evidence is inconclusive.

The origin of the 'grey-gneisses' and their less deformed and metamorphosed equivalents preserved throughout both Early and Late Archean cratons (see below) is most frequently related to some form of subduction processes. The most compelling evidence comes from combined field, petrologic, geochemical and experimental work, which shows that these rocks are intermediates in a multistage evolution from mantle peridotite to Archean granitic crust (Barker, 1979; Barker and Arth, 1979; Anhaeusser and Robb, 1980; Robb, 1983a; Martin, 1994; Ireland et al., 1994; van der Laan and Wyllie, 1992; Wyllie et al., 1997; Rapp, 1995, 1997; Rapp and Shimuzu, 1996; Johannes and Holtz, 1996).

The main arguments for such a plate tectonic boundary-zone setting these granitoids lie in their mineralogy and chemical composition. First and foremost, these rocks are predominantly sodium-rich granitoids with relatively high Al_2O_3 contents (the Tonalite-Trondhjemite-Granodiorite series, or TTG), and secondly they have specific trace element compositions, in particular, relatively high LREE/HREE ratios with marked depletions in HREE (Martin, 1994). When taken in unison these geochemical fingerprints indicate that these rocks formed by partial melting of mafic sources over a restricted range of pressures/depths (7–32 kbars/20–100 km) and temperatures (750–1000°C), mostly in the presence of garnet and hornblende and with or without free water (Fig. 2). Hydrated basalt is an often quoted possible source, and a frequent starting composition in many experiments during which TTG melts are generated (Rapp, 1995; Rapp and Watson, 1995; Johannes and Holtz, 1996; Winther, 1996). Recently mafic granulites have also been used successfully (Springer and Seck, 1997). Isotopic data of TTG indicate that their inferred sources must have been relatively young and that these, in turn, were derived from already significantly depleted mantle sources. The simplest way to achieve this efficiently, so as to yield the great aerial extents of TTG observed in Archean terrains, is through direct melting of hydrated oceanic crust or hydrated gabbros beneath arcs or oceanic plateaux. Other models, however, have also been presented (Kramers, 1987; Ridley and Kramers, 1990; Arndt and Chauvel, 1991; Evans and Hanson, 1997).

The experimental work indicates that 20–40% TTG melt-generation results from dehydration melting near and just beyond the amphibole-out phase boundary, and that the temperatures needed to complete this (1000–1150°C) require relatively elevated geotherms: hot and relatively young, hydrothermally altered basalts would do the trick (Fig. 2).

Very young oceanic crust, for example, formed at spreading ridges close to low angle subduction zones, provides a source and tectonic environment in which modern magmas with TTG characteristics can be produced (Delong et al., 1979; Defant and Drummond, 1990; Drummond and Defant, 1990; Martin, 1986, 1993, 1994; Peacock, 1993; Rapp, 1995; Tatsumi and Eggins, 1995; Karsten et al., 1996). Such modern equivalents of Archean rock suites (Adakites/sodium granites) are volumetrically rare in modern arc terrains, almost certainly because the required thermal conditions in active subduction zones are rare and short-lived (Peacock, 1993). Today, oceanic crust generally subducts at angles too steep to induce hydrous slab-melting; instead, the oceanic crust dehydrates before the metabasalt solidus is intersected. Water-rich fluids are thus extensively generated in modern subduction environments. In turn, these fluids metasomatize and induce melting in the overlying mantle wedge/lithosphere, forming high-magnesium andesitic and basaltic magmas which, with further fractionation, give rise to the andesitic-arc magmatism and associated juvenile granitoids (Baker et al., 1994). Minor low-Al_2O_3 TTG, with subtly different geochemical signatures compared with those of the majority of Archean grey gneisses

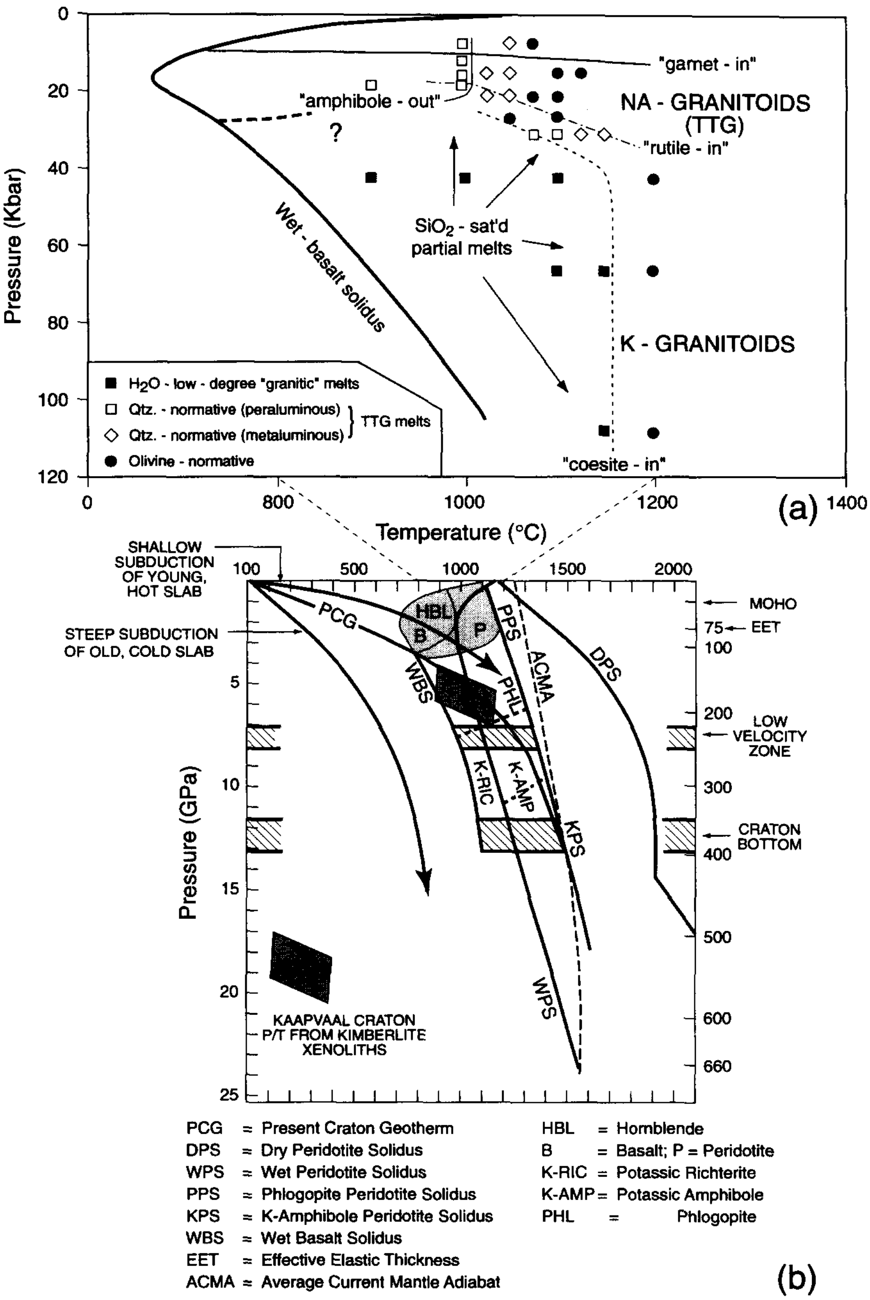


Fig. 2. Simplified temperature–pressure diagrams relevant to the genesis of the crust and mantle keels of Archean cratons.(a). Phase relationships with the water-saturated basalt solidus and the dehydration solidus for amphibole (modified after R. Rapp; pers. communication, 1996). Note the general fields of Tonalite Trondhjemite Granitoids (TTG) and the deeper derived small volume K-granites. Note that at low pressures, the coesite–in curve becomes quartz–in (not shown for clarity).(b). Phase relations showing the fields of Na– and K– amphiboles (including the Ca-rich K-Richterite) relative to the wet basalt and wet peridotite solidus (simplified from Thompson, 1992). Note that shallow subduction of hot oceanic slabs (young oceanic lithosphere) result in dehydration melting at shallow depth to yield TTG (including sanukitoids) up to about 100–125 km, whilst subduction of cold oceanic lithosphere may remove excess water back into the mantle. Also shown are the positions of the internal low velocity zone of the tectosphere and the bottom of the Kaapvaal craton as determined from seismic studies. It is possible that the present bottom of this craton coincides with the intersection of the K-amphibole solidus and the average current mantle adiabat at about 1300–1450°C. (see text for further discussion).

also occur in modern arcs. These are derived through partial melting at lower pressures in the presence of plagioclase, perhaps from partial melting of mafic crust beneath continental arcs (Atherton and Petford, 1993; Martin, 1993, Martin, 1994; Springer and Seck, 1997); but this requires an additional input of water during the melting of the mafic crust.

The above arguments have been used to suggest that Archean subduction, particularly in the Early Archean may have been almost exclusively shallow and at low angles (e.g. Martin, 1994). Independent arguments for such a style of low-angle subduction have been based on models of Archean heat-loss (Bickle, 1978; Abbott and Hoffmann, 1984; Pollack, 1997). It has been further argued that this type of subduction may be the reason for the perceived predominance of bimodal dacite-basalt volcanicity in the Archean, rather than the andesitic volcanicity recorded in the Phanerozoic above high-angle subduction zones (Abbott and Hoffmann, 1984; Harris et al., 1993). However, andesitic rocks in Archean sequences are by no means unknown (de Wit and Ashwal, 1995, 1997); and their apparent scarcity may be due to non-recognition because of severe alteration and or erosion of many Archean terrestrial volcanic sequences (Sylvester et al., 1997a,b).

A distinctly different suite of Archean Mg-rich TTG has been recognized in the Superior Province (Shirey and Hanson, 1984; Stern et al., 1989; Stern and Hanson, 1991; Feng and Kerrich, 1992; Corfu and Stott, 1996; Evans and Hanson, 1997). This suite of rocks has been termed sanukitoid suite, because of their resemblance to modern high-magnesian andesite (HMA) volcanics found in arcs of supra-subduction zones. These rocks may constitute up to 25% of the TTG (Evans and Hanson, 1997). Sanukitoids and HMA (with high Ni, Cr, and LILE enrichments) may be the products of partial melting of a mantle that has been metasomatized by small volumes of TTG-like melt (Rapp and Shimuzu, 1996; R. Rapp, pers. communication, 1996). Thus, TTG and the sanukitoid suite may be continuous with each other, and both the products of slab melting, either directly (TTG) or indirectly (sanukitoid). In terms of their primitive characteristics, sanukitoids and HMA can also be considered part of the compositional continuum of the boninite magma series (Rapp and Shimuzu, 1996); but the trace elements of boninites suggest derivation of partial melting of mantle metasomatized by a slab-derived aqueous fluid (R. Rapp, pers. communication, 1996). This interpretation is in agreement with experimental work which indicates that high to low-Ca boninites can be best interpreted as primary magmas (with 1–8% H_2O) generated by partial melting of fluid-altered (LILE enriched/metasomatized) refractory peridotite at low pressures (3–10 kb), but relatively high temperatures (1150–1350°C), and thus MORB-like geotherms (Crawford et al., 1989; van der Laan et al., 1989; Tatsumi and Eggins, 1995; Evans and Hanson, 1997). Shallow subduction beneath an active or dying oceanic-ridge system would create plausible petrogenetic scenario (van der Laan et al., 1989).

Within a plate tectonic framework, therefore, Archean TTGs might be derived from subducted wet mafic crustal-slabs, but the sanukitoid and boninite sources are mantle-wedge peridotites invaded by low-degree silica- and aqueous-fluids, respectively; and all represent 'hot' subduction zone environments (van der Laan et al., 1989; Rapp and Shimuzu, 1996). From this it follows that even in the Late Archean, relatively steep subduction with overlying mantle wedges may have existed. This may be true too for the Early Archean: in the greenstone belt fragments exposed along the southernmost edge of the Kaapvaal craton, for example, Early Archean mafic rocks (3.2–3.3 Ga) are overwhelmingly HMAs, basalts and diorites with boninitic affinities (Wilson and Versveld, 1994; Riganti and Wilson, 1995; Riganti, 1996). In the Barberton greenstone belt, the basaltic komatiites with pyroxene spinifex, may also have boninitic characteristics (work in progress). These examples suggests that, at least in some places, subduction zones with a refractory mantle wedge between the subducting slab and overlying lithosphere existed in the Early Archean and that partial melting within the mafic crust of (continental margin) arcs may also have contributed to the TTG of the Archean cratons (cf. Pitcher, 1993; Springer and Seck, 1997).

In summary, the evidence from the Archean

TTG suites is consistent with the presence of subduction-related magmatism during the Early Archean, but the data cannot yet demand a verdict on the geometry and style of this subduction process. In terms of some of the sanukitoids of the Superior Province, a caveat must be added: many of the plutons intrude late in the tectonic history of this craton (Evans and Hanson, 1997), apparently well after compressional accretion processes had terminated. These rocks, therefore, are more likely to have been derived directly from late fluid-induced metasomatism of the lithospheric mantle of the craton, perhaps following orogenic collapse (Kusky, 1993). We will return to this later.

Large areas of many Archean cratons are dominated by more evolved and more potassium-rich granitoids (part of the calc-alkaline Granodiorite-Granite-Monzogranite [GGM] suites, including syenites) with I- and S-type granite characteristics (Sylvester, 1994). These intrusives and related extrusives generally postdate the TTG and occur late in the tectonic history of the specific cratons in which they occur. On the Kaapvaal craton, for example, relatively undeformed (3.0–3.1 Ga) GGM-granites can be seen to overly the 3.2–3.6 Ga TTGs as thick horizontal sheets of batholitic proportions, fed by vertical dikes cutting the TTG (Hunter, 1974; de Wit et al., 1992). On other cratons, such as the Early Archean Pilbara Craton and the Late Archean Yilgarn and the Zimbabwean Cratons, the GGM series are far more voluminous than the TTG at presently exposed levels. On these cratons, however, the GGM are more clearly rooted in TTG remnants, and often have round to oval-shaped batholithic dimensions of tens to hundreds of kilometres across (Fig. 3). The notion held until recently that GGM did not exist in the earliest Archean (Taylor and McLennan, 1985, 1995) is no longer tenable (Bowring and Housh, 1995; Kinny and Nutman, 1996; Nutman et al., 1996).

On the strength of the evidence from field and laboratory studies, there is almost complete agreement that the GGM granites were derived through partial melting of preexisting TTG crust and associated sediments at mid-lower crustal levels (Hart et al., 1981; Robb, 1983b; Rutter and Wyllie, 1988; Skjerlie and Johnston, 1994, 1996; Skjerlie et al., 1993; Sylvester, 1994; Bédard, 1996). There is, however, less agreement about the tectonic setting in which this 'continental' melting occurred. In some cases this happened on a craton-wide scale within narrow time spans and without indications of any obvious across-craton chemical asymmetry (Western Pilbara and Eastern Kaapvaal Cratons; Zimbabwean Craton). Elsewhere the magmatism was more locally focused and confined to belts (Superior Province, Yilgarn Craton). Comparisons have been made with the voluminous granites formed in Phanerozoic mobile belts, where these granites (i.e. the Hercynian granites) are believed to have formed in extensional environments following major collision processes (Sylvester, 1994); or wide continental, back-arc rifting regions (Jelsma et al., 1996) such as documented in Southern South America during the Jurassic (Bruhn et al., 1978) or in Japan/China during the Cretaceous (Tatsumi and Eggins, 1995); or within mature arc terrains (Sylvester, 1994). Suffice it to say that the known conditions of the environments must have involved significant increases in crustal thermal-gradients to melt pre-existing continental crust; this could have happened in a number of intracontinental tectonic environments consistent with plate tectonics. Without more supportive evidence from associated rock sequences and structures, GGMs alone, however, provide no robust arguments for plate tectonics in the Early Archean. For this we must turn to the evidence provided by Early Archean greenstones.

4. Early Archean tectonism: evidence from the Kaapvaal and Pilbara greenstones

Only two sizable fragments of relatively pristine Early Archean continental lithosphere are preserved on Earth; they are found in Australia (the Pilbara craton; about 60 000 km^2; Fig. 3) and South Africa (the southeastern sector of the Kaapvaal Craton; about 720 000 km^2; Fig. 4). These cratons have a similar range of tectonic events and litho-stratigraphic sequences dated between 3.0 and 3.6 Ga (Table 2). As both cratons are overlain by Late Archean sedimentary and

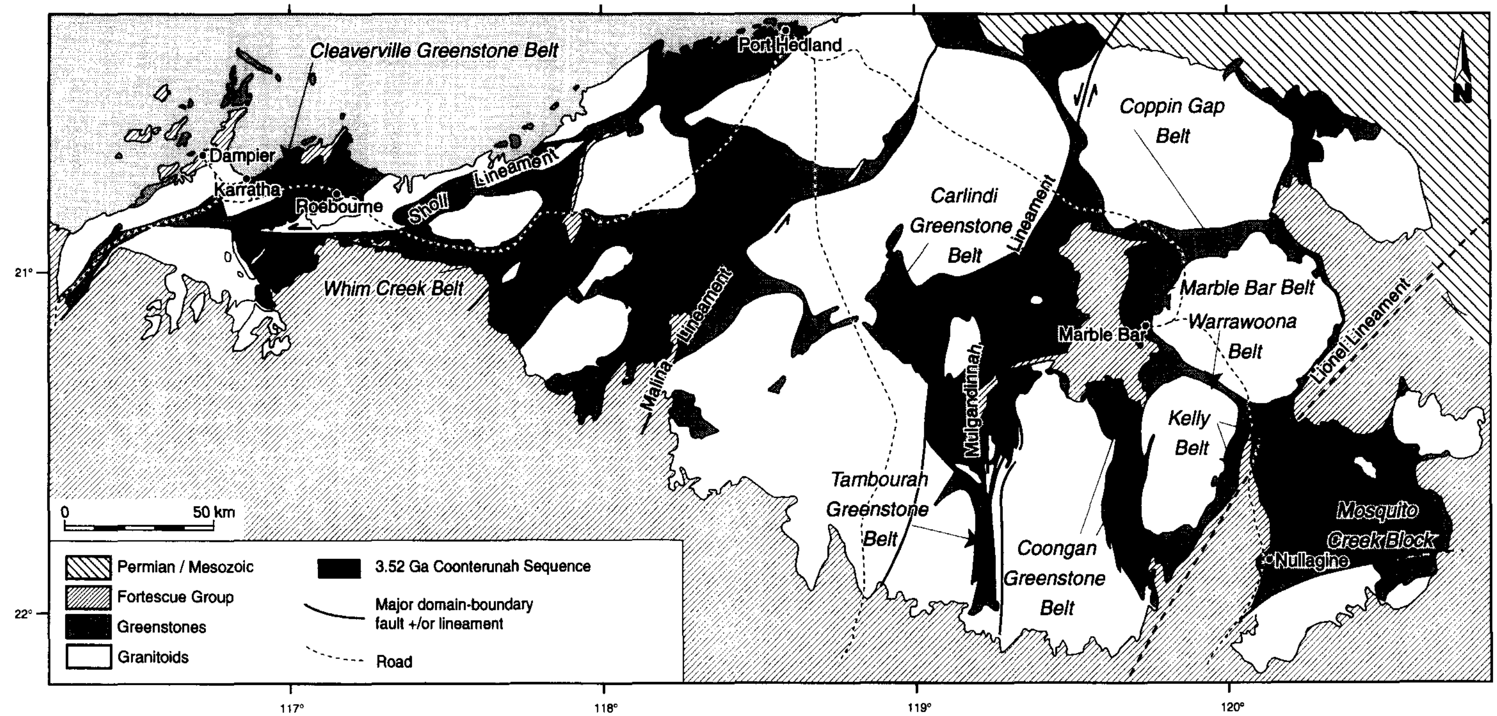


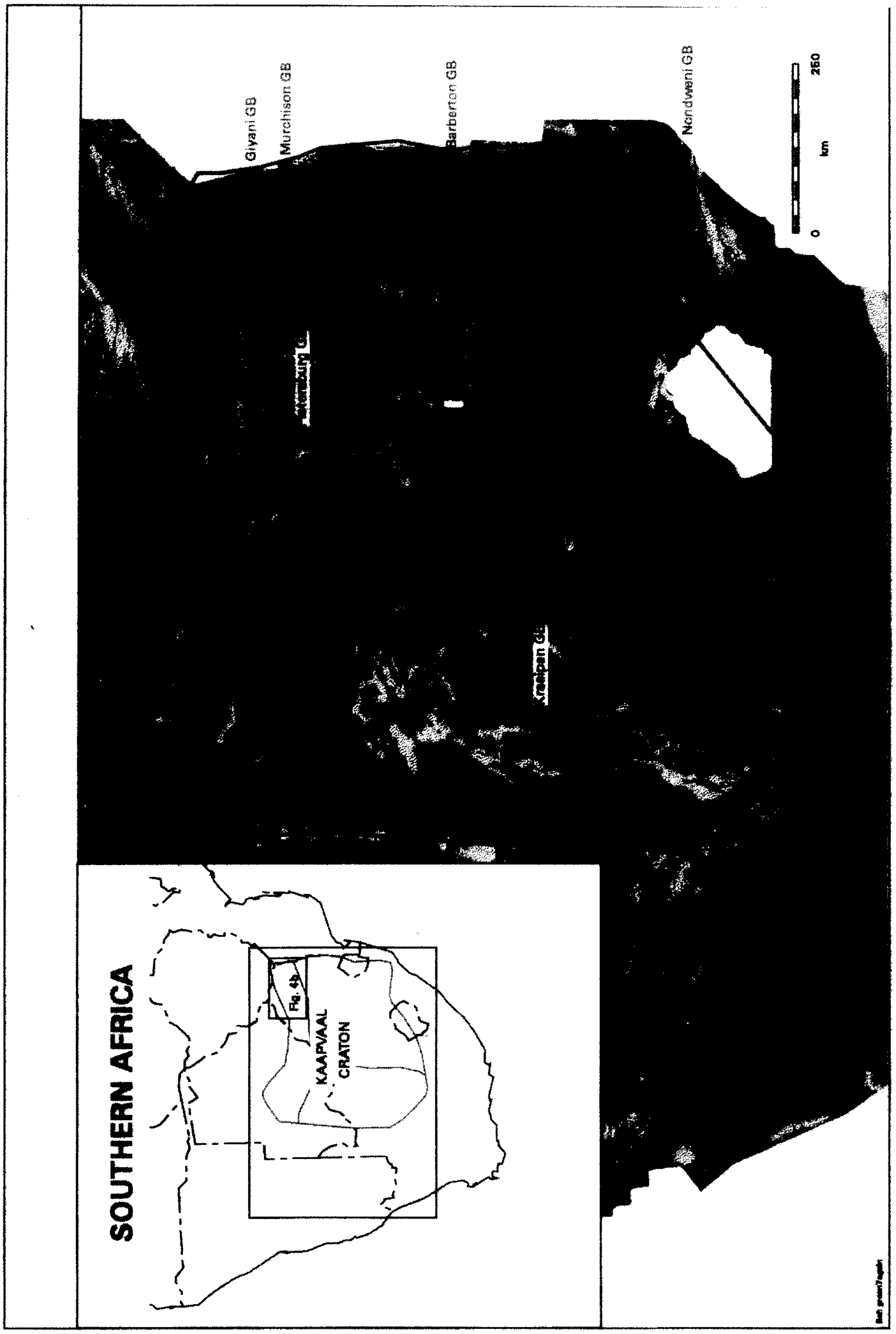
Fig. 3. Regional geological map of the Early Archean Pilbara Craton, NW Australia, showing the typical round to oval shaped granitic complexes (batholiths) and the intervening greenstone belts. Greenstone Belts (GB) referred to in the text are named. The granitoids of this craton comprise an amalgamation of early tonalites (TTG) and late granites (GGM); the latter are clearly rooted in the former. There is at present no satisfactory explanation for the rounded geometric pattern of the batholiths in this and other cratons (e.g. Zimbabwe). Zegers (1996) has clearly shown, however, that the pattern cannot be attributed to deformation induced by simple diapirism as previously suggested by Collins (1989). Structural workby workers of the University of Utrecht (see Zegers et al., 1996; White et al., 1998) suggests that this pattern may be representative of mid-crustal level processes involving extensive in situ melting during extensional tectonics; and that the granite composites represent Archean-type core complexes (also M van Kranendonk, pers. com., 1996. Figure modified after Barley, 1997; T. Zegers and A. Kloppenburg, pers. comm., 1996).

volcano sedimentary sequences of regional extent which, at least in the case of the Kaapvaal Craton, were derived during intense chemical weathering which prevailed across southern Africa, both fragments were parts of stable continental areas by about 3.0 Ga (de Wit et al., 1992; Krapez, 1993, 1996; Trendall, 1995; Fedo et al., 1996; Barley, 1997). Some have argued that these cratons were joined and may have been part of a Late Archean supercontinent, Vaalbara (Cheney and de la Winter, 1995; Cheney, 1996; Zegers et al., 1998a).

The surface geologies of the two cratons have many chronostratigraphic similarities, but their regional surface appearance is quite distinct [Table 2; and compare Barley (1997) with Brandl and de Wit (1997)]. The Pilbara craton, for example, has for long been 'branded' a type-example of a 'gregarious batholith' [McGregor, 1951; Treloar and Blenkinsop, 1995; Hunter and Stow, 1997] in which ovoid composite-granitoids, with ubiquitous gneissose fabrics developed around their margins, dominate the surface structure of the craton (Fig. 3). The Kaapvaal craton lacks this distinctive regional structure (Fig. 4). This may reflect a different level of present surface exposure or it may be due to a different tectonic history.

It has been argued for nearly 50 years that the ovoid to spheroidal structures of granitoid plutons of Archean cratons [of which the Early Archean Pilbara, the Early-Late Archean southern Dharwar (India), and the Late Archean Zimbabwean cratons provide the best examples], represent vertical tectonism driven by regional diapiric granite activity unique to early Earth (McGregor, 1951; Collins, 1989; Ramsay, 1989; Jelsma et al., 1993; Chardon et al., 1996). Others have argued, however, that these structures are interference patterns due to polyphase compressional deformation (see Treloar and Blenkinsop, 1995; Choukroune et al., 1996; Huddleston and Schwertner, 1996; Kusky and Vearncombe, 1997; Ridley et al., 1997, for recent overviews of this debate). Over the last decade or so, some structural geologists have documented detailed observations which appear incompatible with the expected strain regimes of simple diapirs based on experimental work (Schwerdtner, 1984; Huddleston and Schwertner, 1996; Zegers, 1996; Dirks and Jelsma, 1998; van Haaften and White, 1998). However, it is only with the recent development of precise geochronology techniques which can date single minerals formed at a variety of temperatures (U/Pb and Ar/Ar), that the structural and metamorphic history of these granite-gneiss domes are being re-evaluated with the type of rigour needed to resolve this debate. Combined structural and precise geochronology work, particular in the Pilbara and Zimbabwe, has now shown that the diapiric models are gross oversimplifications and of limited applicability to understanding the tectonic history of entire cratons (Bickle et al., 1985; Treloar and Blenkinsop, 1995; Zegers, 1996; Zegers et al., 1996; Dirks and Jelsma, 1998). In the Pilbara region, Zegers (1996) has carefully re-examined the complex kinematic history of the subparallel gneissose foliations around the margins of the ovoid Shaw batholith. This work has revealed a subtle interference, in the ductile environment, between early extensional listric growth-faulting and overprinted low-angle thrusting. Both fault systems are rooted in mid-crustal discontinuities and operated during crustal partial melting to yield a variety of syntectonic granitoid melts which adjusted to the spaces created by the faults. This history is incompatible with a simple diapiric model: the multiple plutons represent what are in essence granitoid core-complexes surrounded by greenstone lithologies.

Other structural re-investigations of shear zones in the Yilgarn, Kaapvaal and the Zimbabwe cratons have also yielded unexpected and surprising results. Williams and Currie (1993), for example, were the first to document extensional loss (of up to 15 km) of greenstone stratigraphy along an Archean ductile shear zone close to a granitoid contact. They inferred from this the cryptic presence of a major Archean metamorphic core-complex. De Ronde and de Wit (1994) similarly interpreted the curved, northern granite-greenstone contact of the Barberton greenstone belt as the edge of a core-complex exposed during ductile extensional faulting and synkinematic granite intrusion at amphibolite facies, following earlier regional thrusting of the greenstones across TTG material at lower metamorphic grades. Re-evaluation of the regional shear zones within and surrounding the greenstone belts of the


Giyani GB
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KAAPVAAL
CRATON

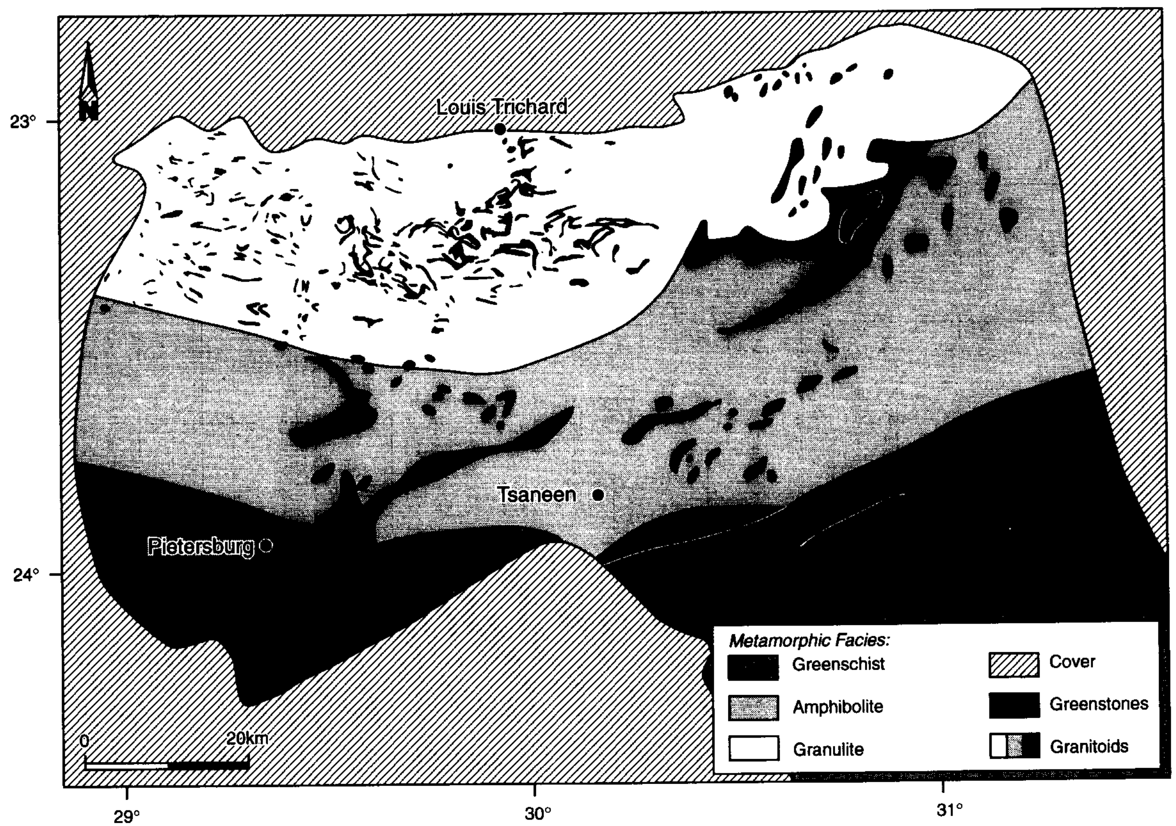
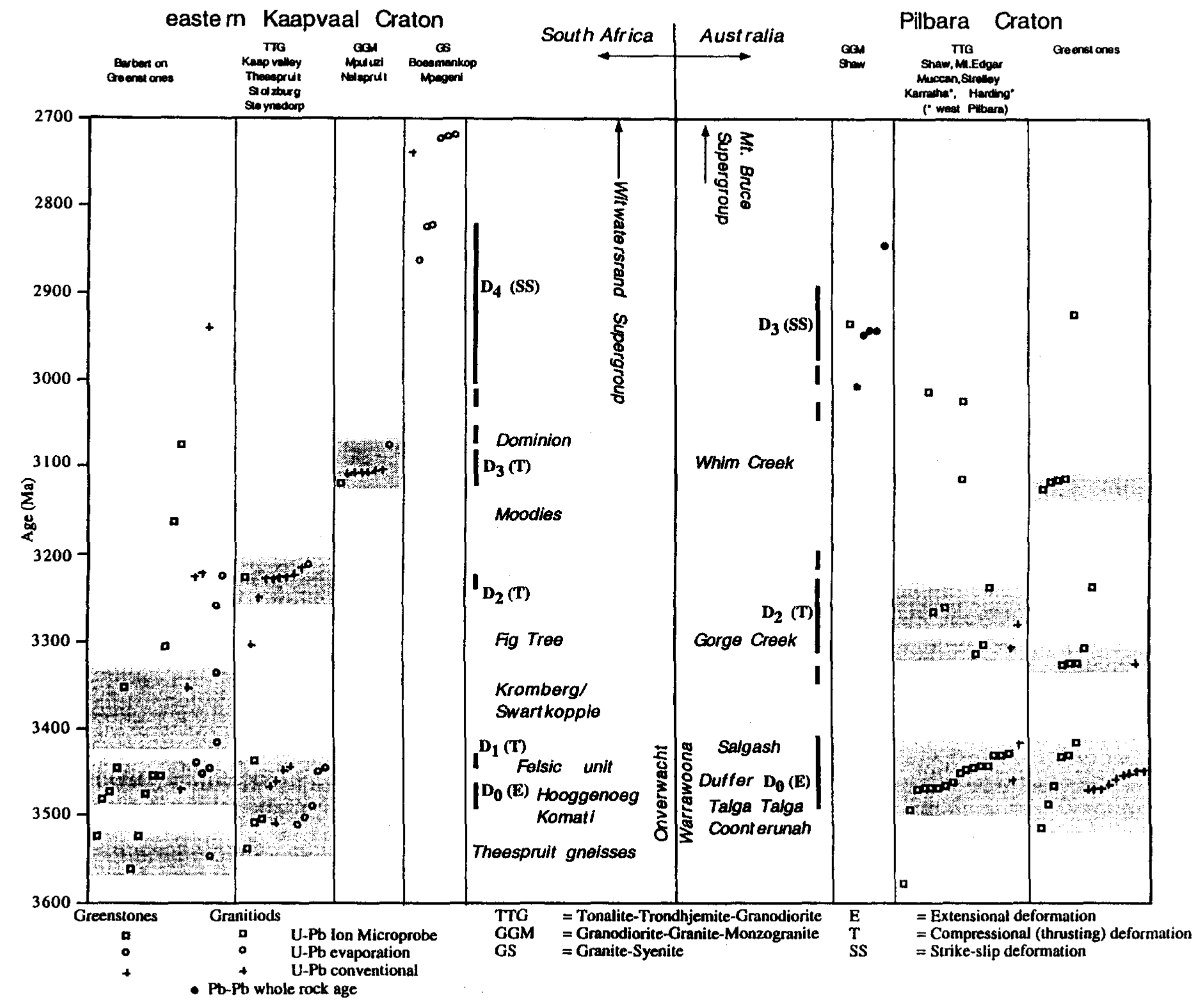


Fig. 4. (a). Aeromagnetic image (grey tones) of the Archean Kaapvaal Craton, southern Africa (from P.B. Leggatt, 1 km-grid total-field map, based on RSA national coverage of the Geological Survey, SA.), with superimposed the major greenstone belts (GB, red). The intervening basement rocks are predominantly TTG and GGM. The granitoids do not display typical rounded batholiths; the GGM are more frequently exposed as large horizontal sheets, suggesting a higher crustal level of exposure than in the Pilbara Craton (Fig. 3). The Kaapvaal Craton can be divided (blue lines) into several major blocks: a predominantly Late Archean domain in the west (Kimberley terrain), and predominantly Early Archean domains in the north and southeast (here informally termed the Southeast and Pietersburg terrains). The Southeast block can be further subdivided into smaller domains (broken blue lines) with distinct tectono-stratigraphic characteristics (de Wit et al., 1992). It is not known if all these small domains are 'terranes' (eg allochthonous with respect to one another). Greenstone belts mentioned in the text are named. In comparing this figure with Fig. 3, it is worth noting that the lateral extent of the greenstone sequences of the (GGM-dominated) Pilbara craton are more continuous than those of the (TTG-dominated) Kaapvaal craton. Geochemical data (Dia et al., 1990) indicate that much greenstone material was eroded from the Kaapvaal craton during the Late Archean. (Figure produced by C. Hatten, Anglo American Research Laboratories, Johannesburg). (b) Distribution of greenstone belt fragments of the southern margin of the Limpopo Belt, within pelitic and TTG gneisses at granulite facies. [For location, see inset on Fig. 4(a)]. It has been suggested that these fragments are high gradeequivalents of the lower-grade greenstone-greenstone terrains of the Kaapvaal craton which were thrust northward across this high grade terrain during the early stages (~2.9 Ga) of amalgamation between the Kaapvaal and Zimbabwean cratons in the Late Archean (de Wit et al., 1993; Treloar and Blenkinsop, 1995; Brandl and de Wit, 1997; Percival et al., 1997). The southern margin is, therefore, the exhumed footwall of a Late Archean regional melange and may be the remnants of a fossil-subduction zone within which TTG were generated, leaving a 'graveyard' of residual mafic and ultramafic pods.

Table 2
Chronostratigraphic comparison between Archean granites and greenstones of the Kaapvaal and Pilbara cratons based on recent U/Pb zircon age determinations. The age-range of various episodes of deformation on both cratons are also shown. This figure illustrates that both regions have a similar geologic history between 3.0 and 3.6 Ga; however there are also some distinct differences which have not yet been analysed in sufficient detail. [Modified from Zegers et al. (1998)].



Zimbabwean craton have similarly produced new kinematic interpretations that do not permit simple granite diapiric models (Treloar and Blenkinsop, 1995). However, all of these studies still lack the careful field-integrated geochronology which sets Zegers (1996) work apart: Zeger has set a new standard against which to measure the complex evolution of the margins between the granites and greenstone belts of all Archean terrains.

Archean diapiric models of granitoid emplacement also face the same general criticism on thermal, mechanical, and rheological grounds, as do Phanerozoic models (Clemens and Mayer, 1992; Treloar and Blenkinsop, 1995; Hutton, 1996). Increasingly, structural studies advocate syntectonic emplacement of granite sheets in contractional and extensional structural traps both in the Phanerozoic and the Proterozoic (Hutton et al., 1990; Grocott et al., 1994; Nédélec et al., 1995; Edwards et al., 1996; Hutton, 1996) as well as in the Archean (Myers, 1976; de Wit et al., 1987b; Zegers et al., 1996; Kusky and Vearncombe, 1997). Such studies are difficult to test, however, in Archean regions which have been overprinted later by bulk horizontal shortening, such as across the Man (West Africa) and the Hebei (China) cratons and, to a lesser extent, the Dharwar craton (Bouhallier et al., 1993; Choukroune et al., 1996); and almost certainly across most of the Early Archean grey-gneiss complexes mentioned in the previous section.

Ovoid patterns of granitoid plutons are also recorded in Proterozoic and Phanerozoic mobile belts (Burke and Kidd, 1976; Lewry and Stauffer, 1997; van Staal and Coleman-Sadd, 1997) and their genesis can be reconciled with core complexes formed within a wide range of modern plate boundary zones (Baldwin et al., 1993); above subduction zones (Andes), within and behind arc environments (western Pacific), across collision zones (Himalayas) or in wider intracontinental settings such as the Basin and Range Province of the Western United States (James and Mortensen, 1992; de Ronde and de Wit, 1994; Treloar and Blenkinsop, 1995).

Suffice it to say that without precise geochronological control on the structural fabrics, none of the structural data described in Archean terrains (e.g. Collins, 1989; Jelsma et al., 1993; Choukroune et al., 1996) need to be interpreted in a framework of unique Archean granitoid tectonism. Recent work in the Himalayas has highlighted the extremely complex history of synkinematic granite plutonism in both contractional and extensional settings. (Edwards et al., 1996).

Both the Kaapvaal and the Pilbara cratons have been subdivided into smaller blocks, bounded by major structural discontinuities (Krapez and Barley, 1987; de Wit et al., 1992). The blocks have distinct tectono-stratigraphic and geochronological histories, but there is not enough data yet to class all of these domains as allochthonous or exotic terranes. As a first approximation, however, it appears that the Kaapvaal craton comprises a number of amalgamated cratonic fragments (de Wit et al., 1992; Fig. 4), and the Pilbara craton is dissected by late transcurrent faults of which at least two are believed to delineate allochtonous blocks (Fig. 3; Tyler et al., 1992; Blake and Barley, 1992; Barley, 1997; Smith et al., 1998; Zegers et al., 1998) Also see van Kranendonk and Collins (1998) for a different interpretation.

Both cratons have five major greenstone belts. Some of these belts, particularly those of the eastern Pilbara, can be subdivided into smaller greenstone belts on the basis of their state of tectonism and/or rock types, but all are essentially structural remnants. For example, the Warrawoona belt, the Marble Bar belt, the Coongan belt, and the Goppin Gap belt are all part of the Marble Bar Domain (Fig. 3; Barley, 1997) and probably contain (parts of) the same lithostratigraphic sequences, which have been subjected to different strain- and metasomatic-histories (Zegers, 1996,Zegers et al., 1996; White et al., 1998; Kloppenburg et al., 1998). Detailed stratigraphic correlations between these belts are not possible without more precise geochronology than presently available. Similarly, part of the Barberton greenstone belt of the Kaapvaal craton has been called the Jamestown schist belt in the past, on the basis of its greater degree of deformation compared to the rest of the belt. Elsewhere on the Kaapvaal craton, greenstone belts separated by tectonic discontinuities may also yet prove to be part of the same original lithostratigraphic sequence (i.e. the Pietersburg and Giyani greenstone belts; Fig. 4). Others, however, clearly comprise similar lithostratigraphic sequences of widely different ages, which have been tectonically juxtaposed, as has been shown for the Barberton greenstone belt (Armstrong et al., 1990; de Wit et al., 1992; de Ronde and de Wit, 1994; Kröner et al., 1996).

Detailed structural field studies of greenstone belts of both cratons have unequivocally documented large-scale horizontal shear zones of both compressional and extensional nature along which regional stratigraphic duplication by thrusting and elimination by listric normal faulting has occurred (Bickle et al., 1980, 1985; de Wit, 1982; de Ronde and de Wit, 1994; Zegers, 1996; Zegers et al., 1996; White et al., 1998; Kusky and Vearncombe, 1997). However, the extent of displacement along any of these shear zones has not been adequately quantified.

In both the Kaapvaal and the Pilbara cratons, the earliest structures recognized are extensional faults. In the Coongan belt of the Pilbara, the extensional tectonics has been clearly identified as contemporaneous with the intrusion of a wedge shaped granodiorite sheet of the Shaw TTG, resulting in features common to core-complexes (Zegers, 1996; Zegers et al., 1996). In the Kaapvaal craton, early extensional tectonics is contemporaneous with the emplacement of extensive mafic-ultramafic complexes (de Wit, 1986; de Ronde and de Wit, 1994; Davies, 1998). In both the Pietersburg and Barberton belts some of these complexes have been identified as being ophiolitic[1], some with MORB-like, and some with arc-like characteristics. The greenstone fragments within the pelitic gneisses of the southern margin of the Limpopo belt [Fig. 4(b)] may also have had a similar origin (de Wit et al., 1987a,b; Jones, 1990; de Wit et al., 1993; Brandl and de Wit, 1997).

In the Pilbara, the possible ophiolitic character of the Talga-Talga sequence has also been recently recognized in the Goppin Gap belt (S. White and J. Wijbrands, pers. comm., 1996). Additionally, in the western Pilbara craton, recent work has documented the presence of a series of duplexes of oceanic-like crust (AMORB or Archean-MORB) within the ca. 3.1 Ga Cleaverville greenstone belt (Otha et al., 1996).

Other mafic–ultramafic complexes in greenstone belts of the Kaapvaal craton have been found to be mainly composed of volcanic arc remnants. The Murchinson belt may have developed into volcanic arc system between 3.0 and 3.2 Ga (Vearncombe, 1991; Brandl et al., 1996), whereas the 3.2–3.3 Ga greenstone fragments along the southern most exposures of the Kaapvaal craton appear to have all the characteristics of young immature arcs, dominated by high Mg-andesites and boninite-like volcanics and hypabbysal intrusions (Wilson and Versveld, 1994; Riganti and Wilson, 1995; Riganti, 1996; A. Wilson, pers. comms, 1996).

In the Barberton greenstone belt, vestiges of volcanic arc rocks associated with TTG-root zones of vastly different ages (ca. 3.45 and ca. 3.23 Ga) are tectonically juxtaposed along a major tectonic boundary marked, in places, by a melange zone. In turn, both these arc terrains are thrust-stacked above and below the ophiolitic-like sequences (de Ronde and de Wit, 1994). In both the Barberton and Pietersburg greenstone belts, integrated structural and geochronological work has unequivocally shown that older greenschist grade mafic-ultramafic sequences were emplaced as thrust sheets across thrust-stacks of younger, tectonically imbricated and kyanite-bearing, felsic-volcanic/volcanoclastic and/or sedimentary (including mature conglomerates) sequences (Armstrong et al., 1990; de Wit and Hart, 1993; de Wit et al., 1993; de Ronde and de Wit, 1994; Kusky and Vearncombe, 1997).

In the eastern Pilbara greenstone assemblages, tholeiitic and calc-alkaline volcanic arc and near-arc environments have also been documented (Barley, 1997). It must be stressed, however, that there is a distinct lack of sufficiently detailed and integrated geochemistry/field-mapping on any of the mafic-ultramafic sequences from greenstone belts of both the cratons, with the exception perhaps of those along the southern margin of the Kaapvaal craton (Wilson and Versveld, 1994; Riganti and Wilson, 1995; Riganti, 1996).

Finally, there is a growing opinion, predominantly by geochemists, that some komatiite-bearing greenstone-sequences may represent oceanic plateaux environments formed at hot spots above deep mantle plumes (Storey et al., 1991; Campbell and Griffiths, 1992; Kusky and Kidd, 1992; Abbott, 1996; Kerr et al., 1996; Kent et al., 1996; Arndt et al., 1997). It may be that within the Early Archean greenstone belts of both the Pilbara and the Kaapvaal craton, vestiges of oceanic plateaux

[1]For an extensive discussion about the use of the term 'ophiolite' in these terrains, compare and contrast Sylvester et al. (1997a,b) with Bickle et al. (1994). Here, I take a similar view as Sylvester et al. (1997a,b): the term ophiolite can be legitimately used for those mafic-ultramafic complexes where ophiolite components have been adequately identified, even if some components, such as sheeted dikes, may not be present or have not (yet?) been identified. Sheeted dikes/sills have recently been found in at least two Early Archean- and three Late Archean-complexes within greenstone belts (de Wit and Ashwal, 1995, 1997; Fripp and Jones, 1997). Renewed field investigations are needed on Archean mafic-ultramafic complexes to reveal the possible range of characteristics of Archean ophiolites.

are also represented. However, at this point two caveats must be added: first, komatiites in greenstone belts of both cratons are minor components (Fig. 5: $<5\%$ of the volcanics of the East Pilbara greenstone belts, and less than 8% and 2% of the volcanics of the Barberton, and the Petersburg/Murchinson greenstone belts, respectively, are komatiites). Only the Nondweni greenstone belt has komatiites in excess of 15% (Fig. 5), but these are mostly pyroxenites ($MgO = 15–22$ wt%) with boninitic affinities, and these are, therefore, probably arc related. Second, the sequence of emplacement in these greenstone belts (and most others) is not as simple as has been proposed for oceanic plateaux (Abbott, 1996): the long held dictum in which ultramafic lavas (komatiites) always dominate the lower strata of greenstone belts, whilst progressively more evolved rocks occur to the top of the greenstone sequences, is now discredited dogma [see Sylvester et al. (1997a,b) for a discussion on this].

From sedimentological studies, there is a substantial body of data that indicates shallow water sedimentation associated with much of the mafic-ultramafic rocks found in the Early Archean greenstone belts (Lowe, 1994; Eriksson et al., 1994, 1997; Nijman et al., 1998a). A recent study of the near perfectly preserved North Pole sequence of the Eastern Pilbara has shown, for example, that this stromatolite-bearing sequence is associated with development of syndepositional barite mounts, dikes and veins flanking normal listric growth-faults (Nijman et al., 1998a,b). Reconstruction of the environment here, as in Barberton, pictures a shallow-water terrain building out from a mafic-ultramafic subaerial platform, perhaps like the Afar or Iceland, dominated by near surface MOR-like processes, including extensive hydrothermal alteration and black/white smoker-like activities (de Ronde and de Wit, 1994; Vearncombe et al., 1995; de Ronde et al., 1997). This is consistent with present-day observations that lavas erupted at shallow ridges are derived from a greater mean depth of melting (and thus greater mean MgO content) than those at deep ridges (Bourdon et al., 1996).

Deeper water sedimentation, particularly Early Archean turbidites has also been recorded on both cratons (Stanistreet et al., 1981; Otha et al., 1996), but there is as yet no study which has quantified the depth of deposition of these sequences, beyond 'below wave-base'.

On both cratons, a wide variety of basins have been documented in fault contact with the mafic-ultramafic sequences: foreland, forearc, backarc, strike-slip and piggy-back-basins, all with striking lithostratigraphic similarity to Phanerozoic examples (Eriksson et al., 1994, 1997). However, there is as yet no sign of a well-preserved sequence of continental shelf-like sediments older than 3.0 Ga, bar the extensive 3.2 Ga metasediments found at granulite facies in the deformed grey-gneisses of the Limpopo belt (Eriksson et al., 1988). Where sedimentation and deformation processes and rates in greenstone basin deposits have been estimated, they compare closely to those measured in active Quaternary strike-slip zones (Lamb, 1986). These studies point to strong lithosphere and Phanerozoic-like crustal flexural rigidities by 3.4 Ga. The presence of substantial strike–slip faults formed between 3.2 and 3.1 within the Barberton greenstone belt and probably throughout the Kaapvaal craton (de Wit et al., 1992; Good and de Wit, 1997), as well as throughout the Pilbara craton (Barley, 1997; Smith et al., 1998), also requires the sort of rigidity of modern plates (Sleep, 1992).

Where modern metamorphic studies have been completed in these and other Archean terrains (and there are relatively few such studies), results yield similar geothermal gradients as found in a wide spectrum of Phanerozoic metamorphic terrains, except those representative of the high pressure-low temperature 'blue schist' facies (Wilkins, 1997).

In summary, there is evidence from the Early Archean greenstone belt sequences that a range of modern plate boundary environments are preserved on both cratons. Moreover, the preservation of AMORB sequences on both cratons implicates melting of shallow upper mantle, welling up only because rigid overlying oceanic lithosphere is actively moving apart. Significant rigidity of Early Archean oceanic parts of plates seems inevitable. As a corollary, there is no reason to believe that plate tectonics did not play a fundamental role in

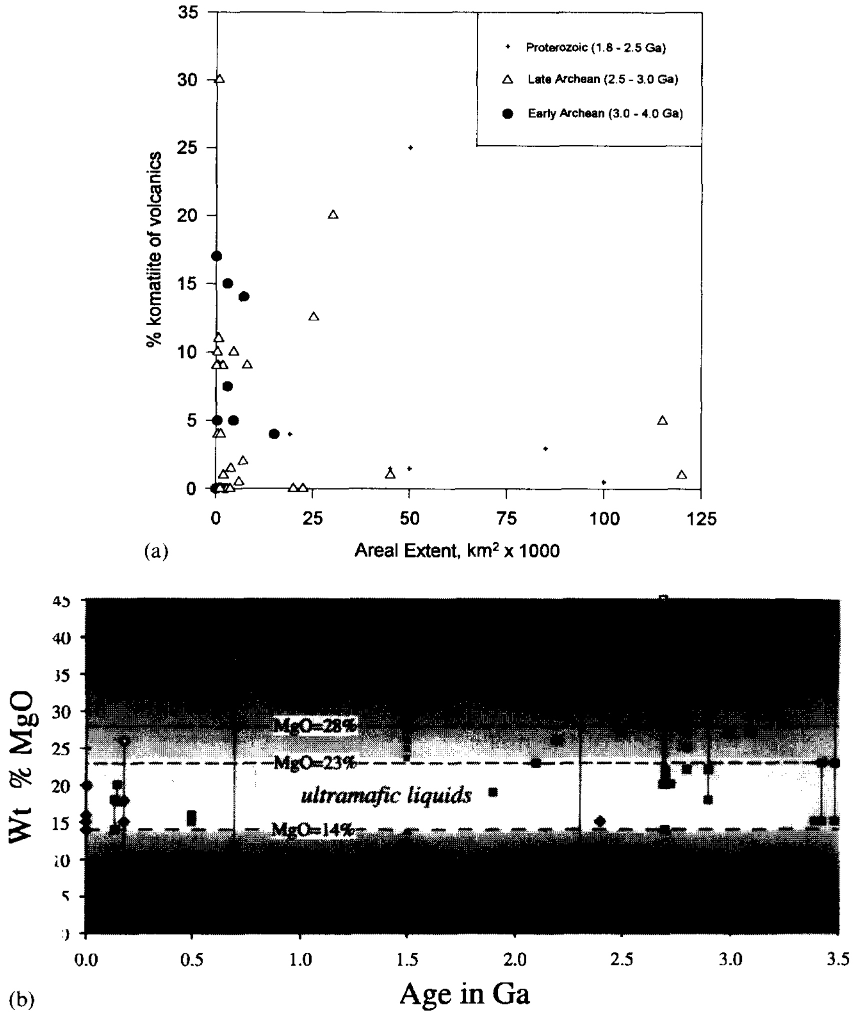


Fig. 5. (a) Areal extent of individual greenstone belts versus estimated per cent komatiite of their volcanic rocks There is no obvious difference in abundance of komatiites from greenstone belts of three selected ages ranges from Paleoproterozoic to Early Archean. Major proportions of komatiites are present mainly in very small greenstone belts (with two exceptions: the Late Archean Carajas, Brazil and the Paleoproterozoic Lapland greenstone belts). Rocks considered as komatiites on this figure range in MgO between 15 and 42% wt % (the later are almost certainly cumulates and the former picrites; data from de Wit and Ashwal, 1995, 1997). (b) Wt% MgO contents of komatiites and picrites plotted against time. Only those compositions varying between circa 14 and 23%—28% are believed to represent primary ultramafic liquids. Open symbols represent compositions of (partial) cumulates and restites; closed symbols are calculated liquidus compositions. A tie line between open and closed symbols represents the known range from rocks of the same greenstone belt, but not necessarily from the same rock suites within the belt. The 23% wt MgO line represent the recently determined maximum liquidus composition of Earths oldest and best preserved komatiites (Parman et al., 1997) from the type section in the Barberton greenstone belt where komatiites were first discovered by Richard and Morris Viljoen (Hunter and Stow, 1997). The 28% wt MgO line represents the maximum liquidus composition of the Archean komatiites as re-calculated by Nisbet et al., 1993. Thus, on this diagram, the maximum decline of ultramafic liquidus compositions from the Archean to the present ranges between 3 and 8 wt% MgO. However, it remains to be firmly established if indeed all komatiite compositions within the 23–28% MgO range represent true liquidus compositions. See text for further discussion. Squares = data from de Wit and Ashwal, 1997; Diamonds = data from other references mentioned in the text; Circles (at 0.0 Ga) represents liquidus composition of recent boninites (Walker and Cameron, 1983); Triangles = Paleoproterozoic bonninite-like komatiites from the Central African Republic (Poidevin, 1993). This figure also clearly emphasis the lack of data available from the Mesoproterozoic (1.0–2.0 Ga).

shaping both these cratons. Field studies of Early Archean plate boundary sequences, however, have not yet revealed anything robust about the sizes, shapes and thicknesses of the plates. Preliminary evidence from the Barberton greenstone belt suggests crust similar in thickness to present-day oceanic-crust, possibly thinner (de Wit et al., 1987a,b). However, theoretical arguments based on inferred Archean mantle temperatures predict much thicker oceanic crust (Bickle, 1986; Vlaar et al., 1994; Arndt et al., 1997). Testing these polarized conclusions is a major challenge for field geologists and experimental petrologists (see below).

Whilst a detailed plate tectonic picture for the Archean is thus still missing, more interestingly, perhaps, is that the 'vestiges of a beginning' of this tectonic process have yet to be found. In both the Pilbara and Kaapvaal cratons, parts of well-preserved sedimentary sequences unconformably overlie small remnants of older deformed continental materials. Within the Barberton greenstone belt, remnants of older TTG have been found (the 3.51–3.55 Ga Theespruit gneisses, Table 2) to be unconformably overlain by the ca. 3.45 Ga felsic sediments of the Onverwacht succession (Armstrong et al., 1990); more substantial sections of such TTG crust are found within the nearby TTG Ancient gneiss complex (Kröner and Tegtmeyer, 1994; Kröner and Byerly, 1995). On the Pilbara craton, Buick et al. (1995) have recently discovered a sliver of deformed greenstones (the ca 3.52 Ga Coonterunah sequence; Fig. 3; Table 2) unconformably overlain by sediments of the 3.45–3.40 Ga Warrawoona greenstone succession. Both these occurrences point to the presence of earlier continental fragments of unknown extent. These then can be linked back in time, albeit still with 50–100 Ma gaps, to the older rock sequences such as the Itsaq gneisses. The latter, in turn provide ample evidence for the presence of stable continental fragments at 3.6 Ga: following 300 Ma of evolution, by 3.6 Ga, the Itsaq Gneiss Complex was able to sustain regional brittle failure as witnessed by a substantial mafic dike swarm (Nutman et al., 1996). There is some fuzzy evidence, therefore, to support models which trace plate tectonic processes as far back as the oldest rocks, the Acasta gneisses.

Some argue that evidence for the onset of plate tectonics will never be found because continental crust may have been forming since the first segregation of Earth's major reservoirs (viz core, mantle, crust), say from about 4.47 Ga onward (see Halliday et al., 1996 and Vervoort et al., 1996 for recent reviews of time constraints), and has merely been recycled more efficiently than at present (Fyfe, 1978; Armstrong, 1981, 1991; Warren, 1989; Bowring and Housh, 1995). The evidence may thus have been effectively destroyed by, for example, extensive meteorite impacts (as modelled on the impact history of the Moon: Green, 1972; Taylor, 1989; Warren, 1989; Goodwin, 1993; Hamilton, 1993; Glikson, 1995). It is also possible, however, that continental crust formation was retarded, and that there is connection between the onset of the formation of regional continents, and the preservation of the earliest continental rocks at 4.0–4.1 Ga (Veizer, 1988; Arndt and Chauvel, 1991; de Wit and Hart, 1993; de Wit and Hynes, 1995). This train of thought, however, requires answering a fundamental question as to what then triggered the late onset of continental crust formation on a large scale; how rapidly did it grow; and how did it affect prevailing Earth tectonic processes?

On the one hand, it has been argued (Arndt and Chauvel, 1991) that the first continent formation (and its effective preservation) was a response to decrease in meteorite bombardment on Earth at about 3.8 Ga. This is not an entirely satisfactory solution because major bombardment is believed to have died away only after the onset of known continental crust formation. It is also not a satisfactory argument because the moon, which had more intense bombardment than the Earth, has extensive 4.3–4.4 Ga rocks preserved. Moreover, the widespread belief that extensive impactite deposits are present within the oldest greenstone sequences of both the Kaapvaal and Pilbara cratons, and which is used in evidence for such late heavy bombardment (Lowe and Byerly, 1986; Lowe et al., 1989; Kyte et al., 1992; Glikson, 1995; Hamilton, 1993; Byerly et al., 1996), has proved to be a false alarm (Koeberl et al., 1993; Koeberl and Reimold, 1995).

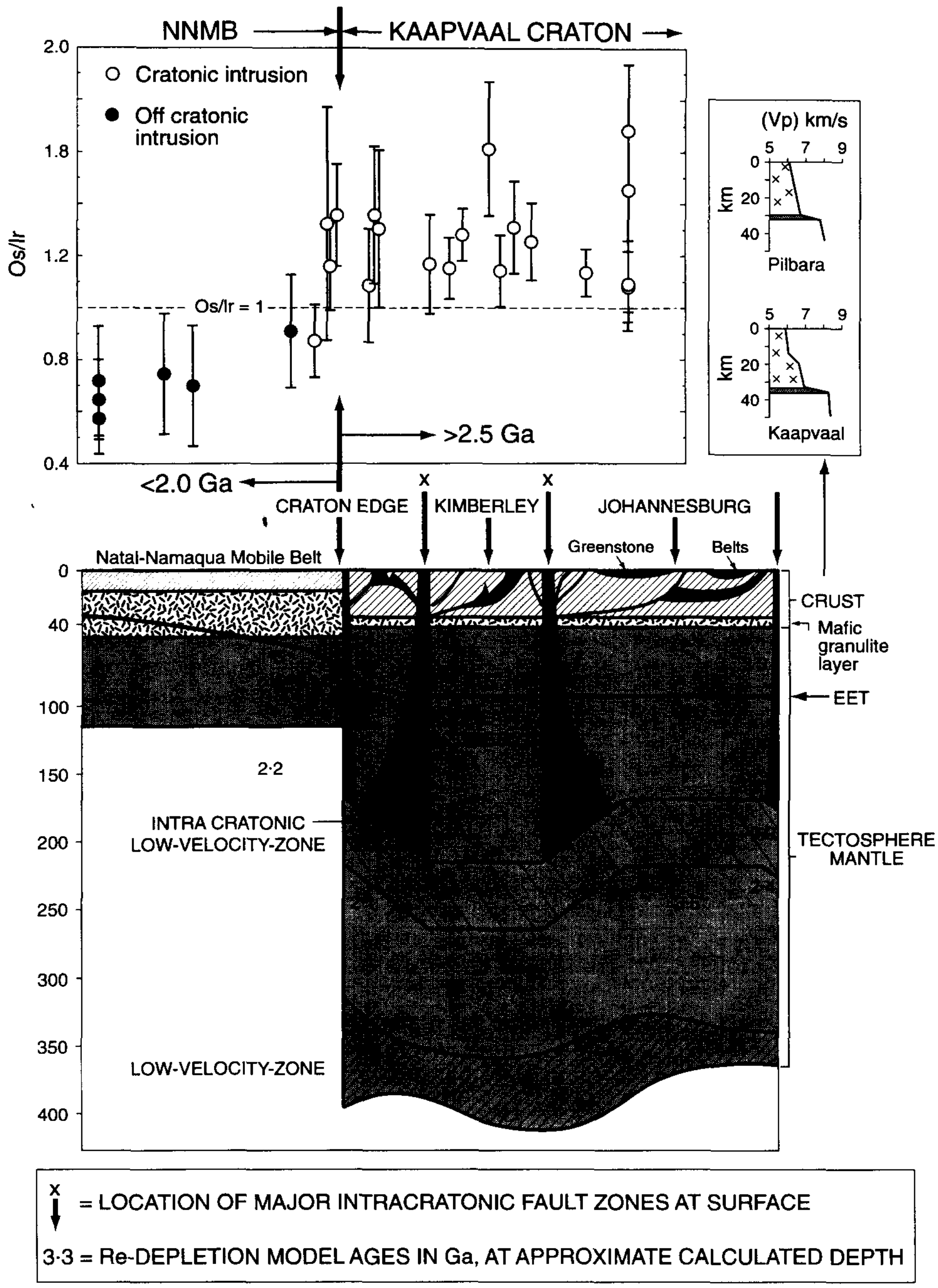
On the other hand, the formation and rapid growth of continental crust has been modelled on the assumption that this coincided with the first time that Earth's MORs subsided below sealevel, thereby triggering the onset of the first large-scale interaction between mafic/ultramafic lithosphere and hydrosphere, (de Wit and Hart, 1993; de Wit and Hynes, 1995). This interaction set off, for the first time in Earth history, efficient production of TTG from hydrated oceanic crust and, in doing so, rapidly created the first continents and moved the Earth permanently into a 'new world' plate tectonic regime at about 4.0–4.1 Ga. Below, I give a brief account of this transition and a possible model for the formation of the earliest continental fragments.

5. Formation of an early Archean craton

Any model for the formation and preservation of Earth's oldest continents such as the Kaapvaal and Pilbara cratons, must incorporate their deep fossil mantle roots or keels (tectosphere of Jordan, 1975, 1988). Recent seismic data give reasons to believe that the mantle root of the Kaapvaal craton comes close to 400 km (Vinnik et al., 1996). Data from mantle xenoliths, brought to surface by kimberlites, as well as limited seismological data have provided only a glimpse of what appears internally to be a very complex keel, with an extended evolutionary history (Fig. 6; Boyd, 1989; Gurney, 1990; de Wit et al., 1992; de Wit and Hart, 1993; McDonald et al., 1995; Pearson et al., 1995a; Carlson et al., 1996; Hart et al., 1997). Several aspects of this keel have now been established to such a degree that they place considerable constraints on tectonic models for their initial formation. Here I list only some of the major ones relevant to this section:

First, Re/Os whole rock data (Pearson et al., 1995a,b) and refractory trace element concentrations in diamonds (Hart et al., 1997) corroborate Sm/Nd model ages of peridotitic diamond inclusions (Richardson et al., 1984), which indicate that a large thickness (>250 km) of the keel was in place by 3.1–3.5 Ga [Fig. 6(b)]. Refractory trace element data on eclogitic diamonds further suggest that at least some of the eclogitic domains of the keel are also of Archean age and that these represent hydrothermally altered Archean oceanic crust (Hart et al., 1997). This is in agreement with radiogenic- and light stable-isotope data on similar eclogites from the Archean Siberian craton (Jacob et al., 1994; Pearson et al., 1995b). Collectively, these observations suggest that simple models of

Fig. 6. Schematic NE–SW section through the Kaapvaal craton based on seismic, gravity, geological and geochemical studies (modified from de Wit and Hart, 1993). Note the distinct different structure and thickness of the cratonic crust compared to that of the adjacent Proterozoic crust; inset shows crustal seismic sections of the two Cratons discussed in the text (from Durrheim and Mooney, 1994).It is known that there are significant local crustal thickness variations of up to 8–10 Km (Muller, 1991). The depth to the bottom of the cratonic keel is taken from Vinnik et al. (1996), and the intracratonic low velocity zone from E. Peberdy (unpublished; see de Wit et al., 1992). The effective elastic thickness (EET) is taken from Doucoure et al. (1996). A distinct seismic anisotropy in the upper lithospheric mantle may be due to old inherited shear zones (but this is not proven and there are other interpretations which suggest that the anisotropy may be due to more recent deformation; Vinnik et al., 1995). The deep shear zones and major flexures to the depth of the low velocity zone may link up with major crustal domain boundaries mapped at surface (de Wit et al., 1992). The inferred increase in width of the shear zones with depth as shown on this figure, is based on field and experimental work on Phanerozoic mantle shear zones and wet olivine rocks respectively (Drury et al., 1991; Vissers et al., 1995). Note that the present thickness of the lithosphere sensu stricto (cf. Anderson, 1995), as defined by the effective elastic thickness (EET), is relatively shallow (about 72 km) compared to ancient cratons worldwide (Doucoure et al., 1996); this Kaapvaal lithosphere has a basal temperature of about 600°C. A similar interpretation has been given to explain both seismic and deep electrical anisotropy within the old, deep lithospheric mantle of the Superior Province (Mareschal et al., 1995; Silver and Kaneshima, 1993). The geochemistry of the cratonic mantle is distinct, both in its major and trace elements as well as minerals and elements of economic significance (diamonds, platinum group elements [PGE], gold). The upper diagram clearly displays this geochemical characteristic in terms Os/Ir ratios (McDonald et al., 1995). The cratonic (tectosphere) mantle also has distinctly heterogeneous domains of Re/Os model ages (selected data taken from Pearson et al., 1995a), which illustrate that the at least some of the lower parts of the keel are of Early Archean age. Such a structure would rule out major episodes of delamination of this cratonic keel. See text for further discussion.


NNMB
KAAPVAAL CRATON
Cratonic intrusion
Off cratonic intrusion
Os/Ir
Os/Ir = 1
>2.5 Ga
<2.0 Ga
(Vp) km/s
km
Pilbara
Kaapvaal
CRATON EDGE
KIMBERLEY
JOHANNESBURG
Greenstone
Belts
Natal-Namaqua Mobile Belt
CRUST
Mafic granulite layer
EET
2·2
INTRA CRATONIC LOW-VELOCITY-ZONE
TECTOSPHERE MANTLE
LOW-VELOCITY-ZONE
= LOCATION OF MAJOR INTRACRATONIC FAULT ZONES AT SURFACE
3·3 = Re-DEPLETION MODEL AGES IN Ga, AT APPROXIMATE CALCULATED DEPTH

keel formation above a one-off Archean deep-mantle plume [Fig. 7(a); Boyd, 1989; Kröner and Layer, 1992; Herzberg, 1993; Haggerty, 1994] are not tenable because such models advocate that the mantle peridotites represent the Early Archean residues of komatiites and basalts; and the latter are known to occur in both the older (3.4–3.5 Ga) and the younger (3.1-3.3 Ga) overlying greenstone belts, spanning a time period of nearly 400×10^6 yr. Moreover, in these models, the crustal (eclogitic) parts (which are now known to have been recycled from oceanic environments, Daniels et al., 1996) of the mantle keel are added later, in post Archean times. Models calling upon the need for separate episodes of tectonic underplating of the different rock types found in the keel, first Archean peridotites and later Proterozoic mafic oceanic crust (Helmsteadt and Schulze, 1989; Gurney, 1990), can now be more rationally extended. Episodic underplating of oceanic lithosphere is still implicated; and the stable isotope geochemistry of the whole rocks and diamonds porphyroblasts indicates that this lithosphere was hydrothermally altered oceanic lithosphere (Schulze, 1986; Gurney, 1990; Daniels et al., 1996; Hart et al., 1997). Major element data (particularly the relatively low Mg/Si and Ca/Al, and the high Mg/Fe), however, indicates that the uppermost (low-temperature) keel peridotites differ from those of modern oceanic peridotites (Boyd, 1989; Boyd et al., 1992). These chemical differences are difficult to interpret: they may represent high-pressure residues formed during extraction of komatiites in a deep anhydrous mantle (Boyd, 1989) and which, thereafter, could have been underplated onto the keels directly from the mantle (Jordan, 1988; Boyd, 1989; Boyd et al., 1992). Alternatively, the peridotites could have been derived at shallower levels from a more hydrous Archean mantle beneath Archean MORs (see below) and underplated tectonically as part of recycling oceanic lithosphere, following extensive oceanic hydrothermal alteration (Schulze, 1986; Hart et al., 1997). In the case of the latter model, loss of secondary volatiles subsequent to tectonic incorporation into the cratonic keel would have further recrystallized, altered (metasomatized) and/or partially melted the peridotites, (cf. Rosing

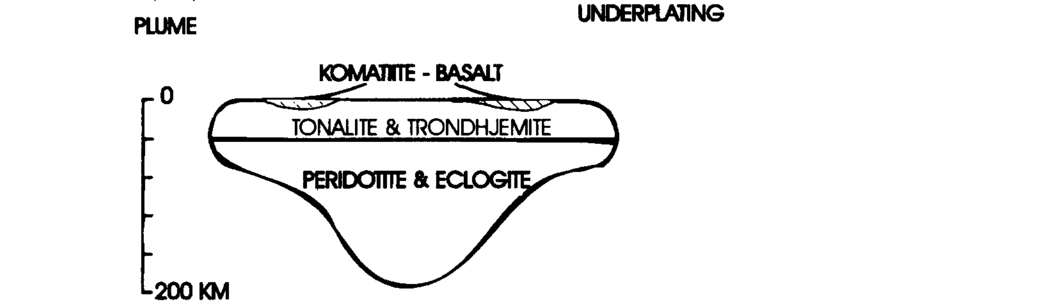


Fig. 7. Cartoon conveying the first order components of the Kaapvaal craton (bottom) and two popular, competing models for the formation of Early Archean cratons and their deep mantle keels (tectosphere): (1) above a mantle plume (left: e.g. Boyd, 1989; Haggerty, 1994); (2) above subducting oceanic lithosphere (right: e.g. Helmsteadt and Schulze, 1989; Gurney, 1990). Some workers advocate hybrid models in which oceanic plateaux have been successively overprinted by marginal subduction processes. See text for further discussion.

and Rose, 1993) to yield compositions potentially matching those of the, by now cratonic peridotites as has similarly been suggested for the eclogites (Ireland et al., 1994).

Second, teleseismic data indicates the presence of a distinct internal low-velocity zone around 200–250 km [Fig. 2(b); E. Peredery, unpublished data; see de Wit et al. (1992); de Wit and Hart (1993)]. Depth to the top of this low velocity zone appears irregular, with sharp gradients beneath major Early Archean crustal shear zones charted at surface of the craton. These crustal shear zones separate different cratonic crustal domains (de Wit et al., 1992). Seismic anisotropy studies (Vinnik et al., 1995) show that along at least one of these shear zones, the Murchinson–Thabazimbi lineament, there is subparallel alignment of olivine in the underlying upper mantle (R. Green pers. comms, 1995; Good and de Wit, 1997).

Third, the mean heat-flow from the Early Archean cratons is low (ca. 40 mW/m^2) compared with the world average continental heat flow; this has been interpreted to be due to the combined effect of low crustal heat production and low mantle heat flow. It has been suggested that the latter is because of the greater insulating effects and the modulation of asthenospheric heat-flow by the cratonic roots (Ballard and Pollack, 1987; Lenardic and Kaula, 1995; Pollack, 1997).

A major conclusion from these data must be that tectonic stacking of hydrothermally altered oceanic-like lithosphere with a high component of depleted mantle, komatiites and basaltic crust, rapidly built up a substantial and thick nucleus of the keel in the Early Archean, and which thereafter remained remarkably stable. As a corollary, major mantle delamination episodes cannot be called upon to explain subsequent Archean (and later) tectonothermal events recorded at the surface of the Kaapvaal craton.

The Kaapvaal and Pilbara cratons have some unique crustal characteristics which are worth summarizing:

First, the crust of both these cratons is thin, but variable (28–42 km; average 35 km) compared to that of Proterozoic crust (40–55 km) and Phanerozoic crust in which no recent extension has occurred (Fig. 6; de Wit et al., 1992; Durrheim and Mooney, 1994). There is, however, considerable variability in both crustal thickness, Moho dip and uppermost mantle velocities (8.27–7.98 km/s) across the Kaapvaal craton (Durrheim, 1989; Muller, 1991; Durrheim and Mooney, 1994). Locally, rapid variations of crustal thickness of up to 8 km over horizontal distances of less than 100 km across parts of the Kaapvaal craton are evident from spectral ratio studies of teleseismic P-waves; these studies have also documented Moho dips of up to 10° with large azimuthal variations (Muller, 1991; R. Green pers. comms, 1995–1997). Changes of 180° in Moho dip-directions suggest regional monoclinal lower-crustal structures (Muller, 1991). The crustal thickness along the northern periphery of the Kaapvaal craton (beneath the high-grade, Late Archean Limpopo mobile belt, where mid- to lower-crustal rocks are exposed) is 40 km thick, and is overlain by denser Moho material than elsewhere (8.4 km/s; Stuart and Zengani, 1987). Within the internal parts of the Kaapvaal Craton, the crust varies between 32 and 42 km, and a two-layer crustal structure is apparent: a mid-crustal P-wave velocity gradient between depth of 8–10 km separates an upper 6.1–6.2 km/s layer from a lower 6.5-6.8 km/s layer (Durrheim, 1989; Durrheim and Mooney, 1994). The transition at the Moho is sharp over 1–3 km, with velocities increasing from 6.8 km/s to subcrustal mantle velocities of ca. 8.2 km/s.

Second, the crust of these cratons lacks a substantial basal high-velocity layer (>7 km/s) of mafic granulites, which characterizes younger terrains (Fig. 6; Durrheim and Mooney, 1994). The overall composition of the preserved Early Archean crust thus appears more felsic in composition than younger crust (Hart et al., 1981, 1990; Durrheim and Mooney, 1994; Rudnick and Fountain, 1995); it is almost as if only a layer similar to that of recent upper continental crust has been preserved. An additional point worth making is that regional seismological studies of the crust of the two cratons may reveal little about their original thicknesses and compositions. Both cratons have experienced almost 1×10^9 yr of relatively active history. During the latter 500×10^6 yr in the Late Archean, episodic basin evolution and GGM formation, must have induced repeated regional extension and contraction during which crustal thickness variations may have been greatly altered. The relatively thin crust of the Kaapvaal craton and its generally felsic nature compared with average continental crust, probably reflects this long period of Late Archean reworking and partial remelting. Seismological evidence alone cannot, therefore, be used to invalidate comparisons of these regions to modern environments as has been done, for example, by Durrheim and Mooney (1994) and Abbott (1996).

Third, geophysical studies have shown that the greenstone belts exposed at surface are essentially shallow sheet-like features, seldom greater than 5–7 km in depth. Their lower boundaries are probably mainly defined by subhorizontal shear zones (de Wit and Ashwal, 1995; Stettler et al., 1997). Geochemical studies suggest that a significantly greater fraction of Early Archean mafic material must have been eroded from the top of the Kaapvaal craton than is presently preserved in its greenstone belts (Dia et al., 1990; de Wit and Hart, 1993). No study of this kind is available from the Pilbara craton.

Keeping the above constraints in mind, the essence of my model is as follows:

Thin oceanic lithosphere of pre-Early Archean (Hadean) times was essentially recycled into the mantle unaltered and dry, perhaps along the lines suggested by Davies (1992) (see also Williams and Pan, 1990). On an Earth with 30% less water stored in its oceans, Hadean oceanic spreading ridges (HMORs) would have stood above sealevel, thereby preventing their efficient hydration as is presently seen at submerged MORs (de Wit and Hynes, 1995). Calculations show that with about 70% ocean water at the surface of an Earth without continents, its MORs would drown below sealevel. If the missing water was stored in Earth's mantle and released during linear Hadean degassing from 4.5 Ga onward, then the onset of this event would have taken place at about 4.0 Ga, given reasonable rates of ocean floor creation (14 km^2a^{-1}) at three times the present-day heat loss and a basal-plate temperature of 1550°C (de Wit and Hynes, 1995). Thereafter, oceanic crust would for the first time have become efficiently hydrated. On such an Earth, the oldest extant ocean floor would have been about 75 Ma (in contrast to about 200 Ma today). In turn, assuming shallow angle subduction of such young and relatively hot lithosphere, this Earth would have had the potential to create TTG-crust at a rate proportional to prevailing spreading rates.

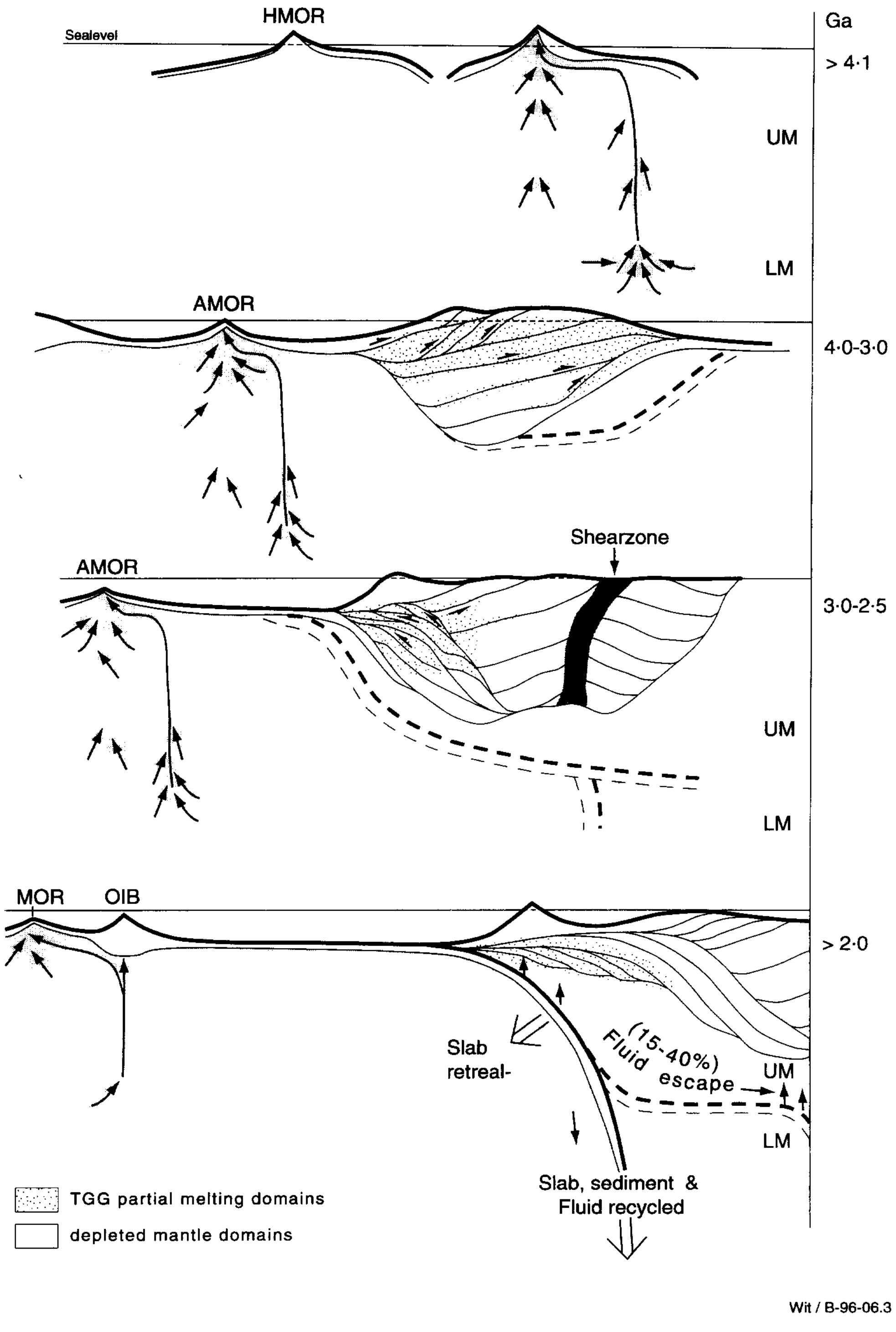
The Early Archean oceanic lithosphere was likely relatively enriched in Mg/Fe, containing a significant proportion of komatiites and underlying depleted mantle perhaps derived from greater mantle depths than today (Tredoux et al., 1989; Nisbet et al., 1993; Herzberg, 1992, 1993; Arndt et al., 1997), although this has not been proved because the depth of the melting column would be dependent greatly on the (unknown) extent of the volatile content of the Archean mantle (Parman et al., 1997; see next section). Hydrothermal flux through the ocean crust at that time would have increased this Mg/Fe ratio further (de Wit and Hart, 1993). Hydration and serpentinization at AMORs would, therefore, have given rise to relatively light (buoyant) upper-oceanic lithosphere. As a result, upper oceanic lithosphere was not amenable to subduction. Slices of young, hydrated oceanic crust and upper mantle restite would have delaminated and formed regional interoceanic duplexes (Fig. 8). The thicknesses of these slices would have been a function of the prevailing oceanic crustal thicknesses and strength profiles (Hoffman and Ranalli, 1988), which would, in turn, have been influenced by the prevailing depths of hydrothermal penetration. Delamination between the upper oceanic lithospheric components from their underlying mantle would have returned to the asthenosphere that portion of the depleted oceanic lithosphere incapable of inducing further dehydration melting.

Here then is the major departure from conventional plate tectonic or hot spot models for continental keel formation (cf. Helmsteadt and Schulze, 1989; Boyd, 1989; Gurney, 1990). In my model, the onset of continent formation is related to an over-stacking process, rather than a more conventional underplating (Fig. 8). Early continents thus formed from the top down. Continuous overstacking with isostatic readjustment would lead to an intraoceanic lithosphere thrust stack, with a thick-

Fig. 8. Cartoon showing the preferred model for the formation of Early Archean cratons as discussed in the text. The earliest fragments of oceanic lithosphere formed at mid ocean-like ridges are thought to have been intensely hydrated and less dense than modern oceanic lithosphere. Such buoyant Archean oceanic lithosphere resisted subduction and delaminated to form stacks of allochtonous sheets (flakes) which upon continuous tectonic-burial underwent dehydration melting to yield tonalitic (TTG) melts between 20 and 100 km (note that at the lowest pressures, TTG formed just above the garnet–in curve, as recorded for trondhjemites of the Dharwar craton; Winther, 1996). TTG melts rose to form overlying and amalgamated crustal sheets. Continued subsidence in turn caused melting of the TTG to yield granites (GGM) at depth of 30 and 35 km, leaving a residue of mafic-intermediate granulites, perhaps during extension due to gravitational collapse during melting at this depth. Hydrated meta-basalts and peridotites continued tectonic burial into the K-richterite and phlogopite fields (see Fig. 2, and text for further discussion).

ness proportional to the density of the hydrated lithosphere. During this tectonic burial process, dehydration melting of the lowermost hydrothermally altered mafic slices would produce TTG. Emplacement of the latter into high crustal levels would have been facilitated along thrusts and extensional shears as recorded in Kaapvaal and the Pilbara cratons (de Wit et al., 1987b; Zegers, 1996, respectively; see Fig. 9) This process would continue as long as overthrusting continued at surface. Any elements not used up in this process would be locked up within this evolving lithospheric keel. For example, K-rich phases such as phlogopite, and K-amphibole and K-bearing pyroxene, would be retained in the deeper buried portions [Fig. 2(b); Thompson, 1992; Foley, 1993; Sato et al., 1997], where they would remain until used up during further dehydration melting to produce small volume alkaline melts, such as K-granites, syenites, lamprophyres, lamproites and kimberlites all of which intruded near-surface, later in the Archean history of the cratonic crust (see, for example, the evolution of the granitoids of the Kaapvaal Craton, shown in Table 2). Tectonic burial would halt once isostatic equilibrium of the now depleted, dehydrated, buoyant and high-solidus keel was attained, perhaps when the bottom of the keel reached the K-amphibole peridotite solidus under the prevailing geothermal gradient (Fig. 2; Thompson, 1992). In this model, therefore, incompatible elements would have been stored in the keel right from its conception, and would not necessarily have to be introduced through later mantle metasomatism. Noble elements, for example, which might have preferentially concentrated in oceanic lithosphere early in Earth history, would also be stored within the keel (e.g. gold and platinum; Groves et al., 1987; Tredoux et al., 1989; McDonald et al., 1995) until tapped by fluids/igneous processes during later tectonothermal events (see for example Groves et al., 1992; Schulze, 1995).

At the upper levels within the thrust pile, TTG-melts would migrate towards the surface, until injected as sheets along thrust- or extensional detachment-zones, to feed volcanic activity at surface, within evolving greenstone belts (Myers, 1976; de Wit et al., 1987b; Zegers, 1996; Zegers et al., 1996). With time and further tectonic burial, the TTG would in turn undergo dehydration-


HMOR
Sealevel
Ga
> 4·1
UM
LM
AMOR
4·0-3·0
Shearzone
AMOR
3·0-2·5
UM
LM
MOR
OIB
> 2·0
Slab
retreal-
(15-40%)
Fluid
escape
UM
LM
Slab, sediment &
Fluid recycled
TGG partial melting domains
depleted mantle domains
Wit / B-96-06.3

melting at 10–20 kbar to yield a great variety of younger GGM melts (Skjerlie and Johnston, 1996); and in doing so finally stabilize an upper crust. Because the process is induced by tectonic burial, melting would have proceeded from the top down. Since the amount of melt that can be produced from a given rock increases with decreasing pressure [because of the solubility of H_2O in silicate liquids increase strongly with pressure, Burnham (1979) and because the stability of amphibole decreases from 10 to 15 kbar, Skjerlie and Johnston, 1996], this downward zone melting would have been very effective in removing GGM melts upwards, perhaps from within an intracrustal extensional regime. In turn, this would have created a sharp gradient with the dense eclogitic/mafic–ultramafic granulite residues. Together with any extensional listric shearing which may have nucleated in the melt zone, these processes may be the underlying mechanisms responsible for developing such a thin Early Archean felsic crust overlying a upper mantle across a relatively sharp Moho.

The above model predicts an evolution of oceanic and continental lithosphere from dry-recycling of oceanic lithosphere without continent formation in the Hadean, to extensive hydration of the oceanic crust and cratonic keel formation in the earliest Archean, followed by shallow subduction, underplating and thick continental lithosphere formation; and finally to the steep subduction, mantle-wedge hydration, trench retreat and thin continental lithosphere production of the present plate tectonic regime (Fig. 9). During the transition from cratonic keel formation to the present style of subduction, slab-pull became increasingly important in driving plate tectonics, whilst flattening of the subducted slab within the transition zone (van der Hilst et al., 1991; van der Hilst, 1995) must have become less prevalent. Similarly, there would have been a concomitant switch from dewatering of the deep mantle and effective near surface retention of water during Early Archean keel formation, to a net return of water (together with a coupled change of fluxes of other elements) to the mantle as a result of recycling within the present subduction regime (Fig. 9; Plank and Langmuir, 1993; Sobolev and Chaussidon, 1996; Staudigel et al., 1998). These different modes of subduction and associated mantle metasomatism probably all contributed to the further shaping of the Kaapvaal keel throughout geologic time, as indeed suggested by the complex radiogenic isotope characteristics of the craton and its long history of subsequent 'anorogenic' igneous intrusions. Because of great potential variations of fluid interactions with the cratonic keels during the evolution of these proposed subduction styles, discourse on subduction and fluid migration beyond the confines of orthodox thinking (e.g.

Fig. 9. Schematic evolution of coupled mid-ocean ridge spreading processes and continental crustal growth, as discussed in the text. The change from 'dry' recycling of mafic-ultramafic lithosphere in Hadean times is linked to the build up of Early Archean cratonic nuclei through the onset of hydrothermal cooling at AMORs. This change resulted from global rise in sealevel due to gradual 'dehydration' of water-rich mantle. By the Early Archean, 70% of the present volume of ocean water was attained, enough to 'drown' AMORs (de Wit and Hynes, 1995). At this stage, 'wet' and buoyant oceanic lithosphere, including sections of depleted upper mantle and oceanic sediments were retained at surface to build up Earths first major continents with thick depleted mantle keels. The tectonic stacking process of the cratonic roots may also have incorporated a composite of stiff refractory buoyant restite materials produced during shallow melt extraction at MORs and deep (off-axis) plumes (cf. Schilling, 1985, 1991; Tredoux et al., 1989; Le Roex et al., 1989; Kaula, 1990; Phipps Morgan, 1997). Materials, hydrothermally-altered at MORs were efficiently used during this stacking process to generate continental crust and were not efficiently recycled to the mantle: continental keels became 'storehouses' for incompatible elements, some of which were removed and/or concentrated into ore deposits during subsequent thermotectonic events. With a change in thermal regime and oceanic lithosphere geochemistry (density), efficient recycling of oceanic lithosphere, and associated sediments and fluids, became possible from Late Archean times onwards. From then on, extensive mantle-wedge melting processes, and backarc basin tectonics became to dominate continental crustal growth processes. In turn this would have altered the mass balance and chemical flux between the crust and mantle. Significant sections of Early Archean continental lithosphere must have foundered back into the mantle to account for (1) material balances apparent in the geochemistry of present day mantle rocks at MORs and OIBs [e.g. Bowring and Housh (1995); Bizzi et al. (1995); de Wit and Hynes (1995), and references therein], and (2) the lack of preserved continental masses of Archean age in the light of projected Archean minimum continental growth rates calculated in this paper.

slab dehydration-melting and asthenospheric-wedge buffering) is a fruitful area for Archean research.

With time, the continental kernels collided to form larger masses. In the case of the Kaapvaal craton this occurred around 3.2–3.3 Ga (de Wit et al., 1992; Fig. 9). These events may have induced Earth's oldest preserved high-pressure/low-temperature metamorphic reactions, now preserved as the diamond facies in the lower portions of the cratonic keel (de Wit and Hart, 1993).

There is a distinct lack of knowledge as to how large these earliest continents might have been; what flexural strengths they may have had; at what rates they may have been displaced; and how these displacements relate to average plate velocities. There are a number of observations that can help resolve these deficiencies in factual information:

The effective elastic thickness (EET) of the Kaapvaal craton is relatively thin at present (72 km) compared to other cratons (Doucoure et al., 1996), but it is not yet known if this was an old, inherited character. Well-constrained sequence stratigraphy from overlying Archean basins will provide further clues in the future.

Palaeomagnetic results from the Archean rock record have been used to argue for fast, slow and stationary states of possible Archean plates (Embleton and Schmidt, 1979; Hale, 1987; Kröner and Layer, 1992). Specifically, palaeomagnetic data have been used to infer that the average Archean horizontal displacement rates of the Kaapvaal and Pilbara cratons between 2.4 and 3.5 Ga were similar to those calculated for todays plate motions and that these two cratons might have moved together as one unit (Layer, 1986; Kröner and Layer, 1992). Although precise dating of the rock sequences on which the palaeomagnetic measurements are conducted has only recently allowed well-dated Archean paleopoles to be determined (Kröner and Layer, 1992), palaeomagnetic signatures from different sites of the same rock sequences (and from the same sites of different rock types) of Kaapvaal craton often give radically different results (Layer, 1986; de Wit and Ashwal, 1997): clearly the ages and styles of remagnitizations, and cryptic corrections for tilting have not been sufficiently tested. The potential for conducting reliable Archean palaeomagnetism has, however, now been amply demonstrated (Hale and Dunlop, 1984; Layer et al., 1989). This must be one of the most important areas in need of development if we are to make meaningful advances in understanding Archean tectonics.

At present judgements must be made, however, on a meagre set of observations; and the Early Archean rock record is sufficiently patchy for us to be deceived by the patterns we see. For example, do the amalgamated fragments that make up the Kaapvaal craton tell us that early continents were small, but deeply rooted continental 'islands', or were these in turn part of earlier more extensive continents with variable lithospheric thicknesses, which have since been broken up and mostly recycled? Implicit answers to these questions are presently mostly based on theoretical arguments about the dynamics of the Earth's Archean mantle. It may be fruitful, therefore, to finish this overview with a few pertinent probes into these theoretical models.

6. Komatiites and a hotter Earth?

Most theoretical arguments about the Archean Earth assume that it sustained a much hotter mantle. These arguments are based on sound knowledge that there was then more radiogenic U, Th and K. Accordingly, radiogenic heat production between 3.0 and 4.0 Ga must have been 2–6 times as high as today; the exact figure depends on the preferred geochemical model of the Earth (Pollack, 1997). There is, however, no agreement as to whether or not the Archean earth had a greater Urey number (a ratio of heat production to heat loss) than today (Williams and Pan, 1992), or whether the Archean Earth may have lost this heat more efficiently than today (for example through a greater global length of ocean spreading ridge, faster spreading and plate motions, thicker oceanic crust production or simply greater devolatilization). Komatiites provide direct observational evidence that has a significant bearing on this problem. The high Mg-komatiites of the Archean are believed to represent very high liquidus temperatures (1575–1675°C, 1 atm.) of melts derived from deep in the mantle (Herzberg, 1992, 1993). Most models, therefore, require mantle sources hotter than the present mantle by up to 350°C. Because komatiites are thought to have been abundant in the Archean, it is generally taken for granted that Earth's mantle in the Archean was much hotter than the present one (see for example Warren, 1984; Nisbet, 1987; Hamilton, 1993; Vlaar et al., 1994). This is inconsistent, however, with other independent estimates (freeboard calculations and observations on the Archean sediment record; and oceanic melt compositions) that the upper mantle liquidus temperatures could not have been much hotter than today (< 100°C; Galer, 1991; Campbell and Griffiths, 1992; Abbott et al., 1994). How can this be?

Wet melts have significantly lower liquidi compared to their dry equivalents; and recent field observations and experimental work has shown that at least some komatiitic melts were hydrous (Grove et al., 1994, 1997; Parman et al., 1997). Allègre (1982) first pointed out that komatiites might be derived from wet magmas similar to boninitic melts found above modern subduction zones (van der Laan et al., 1989). Petrologic and experimental work on Earth's oldest and best preserved komatiites (from the Barberton greenstone belt) suggests that these rocks crystallized from ultramafic melts with MgO contents of 23% and 4–6 wt% of water. Their liquidus temperatures were between 1370 and 1400°C, 200–300°C lower than previous estimates for these same rocks assuming anhydrous komatiite (Green et al., 1975; Nisbet et al., 1993). Thermobarometry using augite indicates that these komatiites were emplaced at depth corresponding to 1.7–2.6 kbar (5–8 km depth), water saturated, as part of a sheeted sill complex (Parman et al., 1997; J. Dann, work in progress). This work shows that extreme high temperatures are not needed to produce the Barberton komatiites. Instead, the komatiites may have been produced in an environment that had a temperature regime similar to that present beneath ocean ridges (primary MORB arrive at the surface at a temperature of 1270–1350°C) and the high water content that characterizes primitive melts above modern subduction zones (Grove et al., 1997). Komatiites may thus represent melts generated above ancient subduction zones with vigorous water activity within the overlying mantle: the komatiite mantle sources are ~200°C higher than the hydrous mantle melting temperatures inferred in modern subduction zones. It has been argued earlier, however, that Early Archean subduction zones may have been shallow and that vigorous recycling of significant volumes of water through the mantle wedge was unlikely.

Alternatively, the komatiites may record the latter stages of degassing of inherited volatiles from the mantle or a hydrous mantle transition zone (Kawamoto et al., 1996; Inoue et al., 1995). Retention of primordal volatiles has been questioned by many workers on the basis that the mantle should have completely devolatilized early in its history. However, the fact that Earth reservoirs with primordal helium are still present today, demonstrates that the Earth has never been completely degassed; and relative abundances of other rare gases (Ne, Xe, Ar) in the present lower mantle are consistent with terrestrial storage of early solar system reservoirs (Honda et al., 1991; Allègre et al., 1993; Porcelli and Wasserburg, 1995; Burnard et al., 1997; Hofmann, 1997a). Thus, retention of significant water in the mantle of the early earth, as required in the model described in this paper to suppress continental-crust formation until ca. 4.0 Ga (de Wit and Hynes, 1995), is consistent with the present data on noble gas isotopes. If the hydrous komatiites were produced in plumes, upwelling from the deep mantle, water must have been more widely distributed than in the present mantle (Kawamoto et al., 1996) since the highest MgO picrites such as these found in Hawaii (~15% MgO) contain less than 0.5% water (Clague et al., 1991). Because there is no present day environment in which komatiite formation can be observed, however, a strict uniformitarian approach to their genesis cannot be applied. It may be equally possible that a hydrous Archean mantle could have produced komatiites over a similar melt column as for basalts at present mid-oceanic ridges and less than 100°C hotter than today (Parman et al., 1997).

The new experimental findings consistent with a wetter Archean mantle, are of fundamental importance in deciding how the Early Archean earth may have operated. Because 'a more hydrous mantle operates at a cooler temperature than an anhydrous one' (Pollack, 1997), the Archean mantle may not have been excessively hotter than the present mantle. The presence of significant dissolved water in komatiitic magmas raises the possibility that the earth lost heat more effectively through dehydration and degassing from a wetter and less viscous and perhaps more vigorously stirring mantle (de Wit and Hynes, 1995; Grove et al., 1997; Pollack, 1997). There is a direct link, for example, between the mantle's viscosity, water content and its thermal evolution (Jackson and Pollack, 1987, Kaula, 1990). Since mantle viscosity is temperature, volatile-content and stress dependant, 'wet' Archean mantle dynamics may have been very different from the present 'dry' conditions: the viscosity of anhydrous mantle, for example, is about two orders of magnitude greater than that of one with only modest amount of water. Modelling this is a difficult task because of the nonlinear and self-regulating relationships between temperature, viscosity and volatiles (Tozer, 1972; Jackson and Pollack, 1987; McGovern and Schubert, 1989; Williams and Pan, 1992).

Nevertheless it is important to continue to evaluate these interactive relationships. For one, they dramatically effect theoretical calculations of the thickness of the Archean oceanic crust and lithosphere. It has been shown that under anhydrous mantle conditions in the Archean, oceans must have had a much thicker crust (30–50 km) than that of the present day (Bickle, 1978, 1986, Warren, 1984, 1989; Sleep and Windley, 1982; Vlaar, 1986; Vlaar et al., 1994; Galer, 1991; Davies, 1992); but this is in conflict with field observations which suggest at least some oceanic-like crust of 3–6 km at 3.5 Ga (de Wit et al., 1987a).

Plume-related hotspot processes are not as an efficient way of losing heat as spreading at mid-ocean ridges. (Sclater et al., 1980; Sleep, 1990; Davies, 1992, 1995; Stein and Stein, 1995; but see Phipps Morgan et al., 1995). It is unlikely, therefore, that the Archean plates were swamped by plumes or replaced by 'plume tectonics' (Davies, 1995) or, more correctly, hotspots (Phipps Morgan et al., 1995). From thermal reasoning, it is not even clear if deep mantle plumes could have existed in the Archean (Sleep et al., 1988). Rather, spreading and hydrothermal cooling processes at the mid-ocean ridges may have been more active (de Wit and Hart, 1993). Field evidence suggests that Archean hydrothermal systems at ocean-like ridges operated more efficiently (de Wit et al., 1987a; Hanor and Duchac, 1990; de Wit and Hart, 1993). This observation has been linked to global hydrothermal heat loss at AMORs of up to an order of magnitude greater at 3.5 Ga than at today's MORs (de Wit and Hart, 1993), implying that the earth's mantle was cooling more rapidly at its ridges than today. This is consistent with findings that near axis hydrothermal cooling at present MORs linearly increases with increasing spreading rates (Baker et al., 1996). Such a scenario would negate the thermal arguments to find alternative processes to plate tectonics for the Archean Earth (Davies, 1995) and would be more 'in tune' with the field observations discussed in this paper.

To summarize, it may well be that the Archean mantle was more volatile rich than today; that MOR processes were more effective in drying and cooling the mantle during the Archean; and that mantle plumes may have been more frequently (and more complexly) focussed through AMORs than they are through MORs today (cf. Schilling, 1985, 1991; Sleep, 1987; Zhang and Tanimoto, 1992; Le Roex et al., 1989; Phipps Morgan and Smith, 1991, Phipps Morgan et al., 1995; Haase et al., 1996). In addition, with time, oceanic lithosphere may have been created through progressively shallower melting at MORs (Williams and Pan, 1992). Such secular changes should still be preserved in the peridotites and the eclogites of the lithospheric mantle keels of the Archean cratons, and could be one of the reasons for the recorded discrepancy between the chemistries of present oceanic peridotites and those found in samples from the present cratonic keels (cf. Boyd, 1989). Finally, there is no firm observational evidence that Archean mantle and plume melts were significantly hotter than those of the present.

7. Secular changes, crustal recycling and the preservation problem

Probably the most contentious debates about Archean tectonics centre around qualitative statements about the relative increase or decrease of certain rock types or rock associations; and relating these to inferred smooth or episodic secular changes of Earth processes. A good example is the often quoted, and generally accepted 'paradigm' of a decrease of komatiite abundance from the Early to Late Archean and in turn to the Proterozoic (Condie, 1994; Nisbet and Walker, 1982, Nisbet, 1987; Ludden et al., 1996, Abbott, 1996). This paradigm is probably the most often used to substantiate a secular decrease in Earth's heat content and/or mantle temperature (Bickle, 1978, 1986, Sleep and Windley, 1982; Richter, 1985; Vlaar, 1986; Vlaar et al., 1994; Nisbet, 1987; Hamilton, 1993; Abbott, 1996; Pollack, 1997). An analysis of the rock make-up of 40 greenstone belts world-wide, for which there is reasonable quantitative data available (out of a potential number of at least 260) indicates, however, that there is no unequivocal decline in the volume of preserved komatiites from the Early Archean to the Phanerozoic (de Wit and Ashwal, 1995, 1997). There is, however, a peak in komatiite preservation in globally distributed greenstone belts of the Late Archean, but this observation must be evaluated in the light of the fact that small greenstone belts (with two Late Archean and two Proterozoic exceptions) have major komatiite volumes [Fig. 5(a)]. There may well have been selective preservation in these tiny greenstone belts.

Today, most workers would agree that liquidus compositions of the most magnesium-rich Archean komatiites range between 23 and 28% MgO; and the best estimates for the highest MgO liquidus contents for Earth's oldest best preserved komatiites (ca. 3.5 Ga, in the Barberton greenstone belt) is 23% (Parman et al., 1997). In the light of this, it is difficult to argue for a substantial secular change in MgO content of ultramafic magmas between the Archean and the Present [Fig. 5(b)]. For example, recent work in the Jurassic (183 Ma) northern Lebombo, Mwenezi and Tulli areas of the Karoo Igneous Province suggests a volume of approximately 40 000 km^3 of picrites and picritic basalts, derived from parental magmas with liquidus composition of 15–18 wt% MgO (MgO contents of the extrusives range between 9 and 26 wt% MgO), within an area comparable to the largest preserved Archean greenstone belts (A. Duncan, pers. comm., 1995-1997); and the Cretaceous (155 Ma) picrites and komatiites on Gorgona Island, with liquidus MgO compositions of 18-20 wt% (Kerr et al., 1996) cover more than 60% of an area comparable to many small Archean greenstone belts. Also, there are present day OIB picrites on Hawaii with 15 wt% MgO liquid compositions (Clague et al., 1991; Anderson, 1995a,b) and 14–16 wt% MgO on Iceland (Elliott et al., 1991). This means that the change with time of the primary MgO composition is less than the scatter at any one time [Fig. 5(b)]. One possibility remains, therefore, that extraction of picritic and komatiitic liquids from the mantle has been, and still is at present, the dominant form of mantle melting (Liang and Elthon, 1990; Anderson, 1995a). A major challenge, therefore, remains a critical assessment of primary MgO compositions in rocks which might have cumulated olivine and/or have undergone MgO addition during hydrothermal alterations (de Wit et al., 1987). For example, in Abbott et al. (1994) there are no plots of MgO versus time, but rather calculated liquidus temperatures versus time; such plots are inappropriate given the low liquidus temperatures of hydrous komatiite melts, and the possible variable degrees of volatites in primitive melts generated at different tectonic settings (Parman et al., 1997).

Similarly, it has been remarked that because there are few mafic alkaline rocks described from Archean terrains, that mantle conditions were not appropriate (too hot) for the formation of this type of magma (Hattori et al., 1996). Blichert-Toft et al. (1996) recognize that this too might be merely an expression of lack of preservation. For example, no Archean kimberlites have yet been discovered from the most ancient Archean terrains. Yet, by proxy, the presence of a significant number of diamonds found in the 2.7–2.8 Ga sediments of the Witwatersrand basin overlying the Early Archean basement of the Kaapvaal craton, illustrates that they must have been present; and by virtue of the fact that so little of the Archean crust may be preserved, one might even argue that there may have been relatively greater numbers of kimberlites present in the Early Archean than in the Phanerozoic. Late Archean lamprophyric dykes in the Limpopo belt also attest to the occurrence of Archean metasomatism and alkaline magmatism rooted in the mantle keel of the Kaapvaal Craton (Watkeys and Armstrong, 1985); and the late shoshonitic alkaline magmatism in the Superior Province (Hattori et al., 1996) may have a similar origin.

Similar argumentation is frequently used in the documentation of sedimentary rocks and mafic volcanic rocks. On the one hand for example, Eriksson and Fedo (1994) emphasize that the preserved stratigraphic records suggest an increase in the volume of synrift and stable-shelf sedimentary successions from 3.2–2.5 Ga '*in response to the growth of stable continents during the Late Archean*' (italics mine, because this part of the statement reflects another common belief that continental growth led to stable continents only during the Late Archean, which has not been proven). On the other hand, Lowe (1994) states that Early Archean greenstones are dominated by volcanic rocks formed during oceanic spreading and/or hot spot volcanism whereas Late Archean belts appear to be dominated by volcanic rocks formed along convergent plate margins.

There are no quantitative data to substantiate these and similar sweeping statements (de Wit and Ashwal, 1995, 1997), but because they are prone to be exploited by theoretical modellers, they need much more careful formulation.

On a global scale, the arguments associated with positive continental crustal growth through time, versus a steady state, or even negative crustal growth with time are of a similar nature. Most continental growth curves are preservation curves, based on the aerial extent of exposed rock sequences of specific ages (Warren, 1989; Bowring and Housh, 1995). Because evidence may not have survived at the surface of the Earth, these problems can only be tackled in conjunction with appropriate geochemistry. To date, researchers have not been successful in solving this thorny 'growth versus recycling' problem satisfactorily (Bowring and Housh, 1995; Hofmann, 1997a,b; Sylvester et al., 1997b). Until they do, models based on apparent uniqueness of rock associations, or the assumption that there is a distinct punctuation in (Archean) continental growth rates (e.g. at 2.7, 3.5, 3.8 etc., Davies, 1995; Stein and Hofmann, 1994) must be treated with caution (as a cursory glance at Table 1 confirms). The 'preservation problem' is an occurring theme which induces circular arguments. Having said that, there is some direct data available which may help to focus further continental crustal growth modelling more clearly:

Given the size of the Superior province and time-span within which 90% of its rocks formed (1.4×10^6 km^2 by 38 km crustal thickness/100 Ma), this section of Late Archean continental crust was created at a net rate of some 0.54 km^{3}y^{-1} (Table 1). If we assume that this rate reflects general minimal Late Archean global crustal growth rates, and if we assume no crustal recycling, then we should expect to find more than 170 Late Archean cratons the size of the Superior province to be preserved on the present planet. This is clearly not the case. Moreover, recent trace element work (Nb and U) on basalts from the Late Archean Yilgarn Craton, suggests that at least at 2.7 Ga, a similar amount of continental crust existed as today (Sylvester et al., 1997b). These conclusions require substantial recycling of continental crust, despite the objections of Hofmann (1997a).

Similarly, it has been calculated that if continental growth rates were 3-4 times the present Phanerozoic rates (1.06 km^{3}y^{-1}; Reymer and Schubert, 1984), then between 3.2 and 3.6 Ga, the worlds continental surface area must have increased between 64 and 85 times the size of the entire Kaapvaal craton (de Wit and Hart, 1993). Where is it now? In the light of only two preserved Early Archean cratons, models postulating substantial recycling of Early Archean continental crust back into the mantle are difficult to avoid (Fyfe, 1978; Armstrong, 1981,Armstrong, 1991; de Wit and Hart, 1993; Bowring and Housh, 1995). Put in a different way, using the minimal net continental crustal growth rates that can be calculated from the known crustal ages and volumes preserved on the Kaapvaal and Superior cratons (0.07 and 0.54 km^{3}y^{-1}, respectively; Table 1) and assuming that Archean growth rates were smooth rather than punctuated [neither of which have been proved; compare, for example, Gurnis and Davies

(1985) with Davies (1995)], then at least 250×10^6 km^2 surface area of Archean crust (more than three times that of the areal extent of continental crust today) should have been preserved by the onset of the Proterozoic. This does not appear to be the case, unless continental crust recycling became more efficient in the later half of Earth history. These back-of-the-envelope calculations vindicate geochemical arguments of Armstrong (1981, 1991) and Bowring and Housh (1995) for very effective processes of continental crust recycling throughout the Archean.

To summarize, the quest of how to figure out a robust way of calculating rates of Archean crustal growth and recycling remains a first order challenge; if only because this will bear on understanding the evolution of all geochemical cycles. At present the best we can say is that it seems intuitively correct to assume that substantial rates of continental crustal recycling occurred as far back as the earliest Archean; and that this is consistent with parametrized models which incorporate predominantly whole mantle convection throughout Earth history (Breuer and Spohn, 1995). Because in the model of craton formation described above recycling by subduction processes was not efficient in the earliest Archean, it seems instead that 'flushing' of proto-cratons may have dominated this recycling process, perhaps in a manner similar to the episodic, non-equilibrium flush-events envisaged for accumulations of oceanic lithosphere at the upper to lower mantle transition zone (e.g. Ringwood, 1991; Hondu et al., 1993; Tatsumi and Eggins, 1995; Yuen et al., 1996).

8. Conclusion

It has been argued that the difference between Archean and Phanerozoic tectonic processes are largely differences of degree rather than style (Dewey and Windley, 1981). Others have argued for plate tectonics (Burke and Kidd, 1976; Burke, 1997; Tarney et al., 1976); whilst others still claim that the evidence for a similar style is lacking altogether. Hamilton (1993), for example, states that 'the products of modern plate tectonics differ so greatly from Archean (rock) assemblages that Archean processes must have been strikingly different from modern ones'. This paper shows that Hamilton's statement is not in accord with the preserved rock record, but are based more on absence of some expected evidence (e.g. melange deposits and blueschist facies metamorphism), rather than on contradictory evidence.

The geologic evidence in favour of plate tectonic processes operating in Late Archean time is solid; and that for Early Archean is more than compelling. Nevertheless, this paper argues that the style of plate tectonic processes at convergent plate boundaries has changed. This is most evident in the Early Archean when recycling of oceanic materials may not have dominated continental crustal-growth in the same style as is evident today. One consequence of the model presented is that back arc spreading may not have been prominent in Early Archean times. It may be that the onset of plate tectonics as we know it today had its roots in the transition time during which Earth's spreading ridges sank below sealevel and when the oceanic lithosphere at spreading centres first started interacting with the hydrosphere, between 4.0 and 4.2 Ga, laying the foundations of Earth's first continents. Before that time, excess water in the mantle may have sustained thin oceanic plate tectonics with efficient plate recycling and rapid cooling of the interior (cf. Williams and Pan, 1990). These conjectures have major consequences for mass balance considerations: a changing plate tectonic Earth from one without recycling of its external hydrosphere and continental materials, to one in which up to 60% or more of its hydrated, metasomatically zoned oceanic lithosphere and other surface components are recycled into the mantle through subduction processes (e.g. Peacock, 1990, 1993; van Bergen et al., 1993; Plank and Langmuir, 1993; Tatsumi and Eggins, 1995; Sobolev and Chaussidon, 1996; for overviews and references).

There is still no detailed tectonic picture of the Early Archean Earth. This paper has outlined some avenues that should be explored further. Particularly in the light of the new era of 'fast' accurate dating and palaeomagnetic measurements, more reliable quantification of geochemical fluxes and tectonic displacements will become known. New observational and experimental work, particularly on komatiites and other ultramafic materials, will shed more light on Archean mantle temperatures and viscosities; and on the structure of oceanic lithosphere. These studies will require international multidisciplinary programmes.

Whilst reiterating the warnings of the geologists that 'things are more complex' and that the detailed kinematics of the tectonic events, even of the well studied Superior Province, are far from understood, much progress has been made; and some of 'missing links' are being found (for melanges, see for example Wang et al., 1996; and for examples of relatively high-P/low-T metamorphism, the diamond facies in the roots of cratons may be considered). But more is needed. Foremost, 'Precambrian' geologists need to realize more than ever before that they must integrate better with those geophysics and geochemistry colleagues who at present are 'locked into' Phanerozoic times. This will require concerted communication efforts about the challenges offered by early Earth rocks and about the fact that 'Precambrian' geologists are ready to come out of their 'Precambrian cocoons'. Canadian geoscientists have set a great example; and the pace.

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