Crust Formation and Plate Motion in the Early Archean

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Mounting evidence for voluminous continental crust formation in the early Archean involving intracrustal melting and selective preservation of granitoid rocks suggests that initial crust formation and growth were predominantly by magmatic underplating in plume-generated Iceland-type settings. Collision of these early islands to give rise to larger blocks is suggested by extensive horizontal shortening in both supracrustal and granitoid assemblages. Preservation of early Archean high-grade gneisses that were once at depths of 20 to 30 kilometers implies that these blocks developed thick, subcrustal roots despite high mantle heat flow. Rigid continental plates must have existed since at least 3.5 billion years ago, and greenstone belts (composed of mixed metavolcanic and metasedimentary sequences intruded by granitoid plutons) probably developed on or near these microcontinents. Paleomagnetic data with good age control from at least one ancient craton suggest that plate motion was at normal minimum average velocities of about 17 millimeters per year with respect to the poles during the period 3.5 billion to 2.4 billion years ago. If this is true on a global scale, Archean plate motion was not faster than in later geologic times.

Discussions on the evolution of the early continental crust during the Archean period 4 billion to 2.5 billion years (Ga) ago are currently dominated by four major issues, namely, (i) the mechanism of earliest crust formation and survival, (ii) crustal growth and differentiation, (iii) the question of whether Archean rock assemblages can be explained in terms of modern-style plate tectonics, and (iv) the rate of motion of continental plates, if any, in the Archean.

Archean terranes of the world (Fig. 1) are characterized by broadly similar rock assemblages and structures (*1*) and are largely composed of three distinct rock units:

1) The most voluminous rock is a gray, heterogeneous, banded gneiss, which, in most cases, has been derived from granitoid precursors of the so-called TTG suite [tonalite-trondhjemite-granodiorite (*2*)]. Some of the best examples of this suite are exposed in West Greenland, in eastern Labrador, Canada, in the Wyoming Province of the United States, in Finland, in Western Australia, in southern India, and in Swaziland, southern Africa, and have been the subject of detailed geochemical and isotopic studies [for recent summaries see (*3–5*)]. Many of these amphibolite- to granulite-grade rocks reveal a complex structural history not seen in adjacent lower grade greenstone belts, and this, together with precise zircon dating (*6–8*), suggests that at least some of them are younger in age than the gneisses. Myers (*9*) developed a scenario in which these original granitoids intrude as sheets into older crust and immediately become involved in intense deformation, culminating in large-scale thrusting and crustal thickening. During this process, many of the granitoids become sheared and attain a banded appearance, which obscures the original contact relationships and may lead to their being mistakenly identified as metasedimentary rocks.

2) Greenstone belts are volumetrically less significant and generally define irregular-shaped, elongate bodies that are almost always in tectonic contact with the gray gneisses discussed above. They consist of varying proportions of predominantly mafic, dark gray to green volcanic rocks (hence the name "greenstone"), tuffs and a variety of metasedimentary rocks including metagraywacke, shale, carbonate rocks, chert, metaquartzite, and banded iron formation (*10*). The volcanic sequence is frequently bimodal, that is, it consists of Mg-rich mafic to ultramafic varieties known as komatiite and komatiitic basalt (*11*) and of minor felsic varieties generally of dacitic to rhyolitic composition (*12*). The komatiites probably reflect large degrees of partial melting in the upper mantle, from which these rocks are derived (*13*), and this mantle was probably 200° to 400°C hotter then than it is today (*14, 15*). The greenstone belts are variably deformed and generally of low metamorphic grade, although some higher grade belts also occur. Greenstone deformation was originally ascribed to a dominance of vertical tectonics (*16*), but recent work suggests strong horizontal shortening and thrusting in most belts. This has led to tectonic stacking and has made it difficult to reconstruct the original stratigraphies (*7, 17–19*). Our ongoing work in the early Archean Barberton greenstone belt in South Africa, one of the best studied and classical greenstone belts of the world (Fig. 1), suggests that the belt may consist of distinct terranes which considerably differ in age and in tectonic history (*20*). A terrane boundary has also been suggested on the basis of isotopic data in the Pilbara craton of Western Australia (*21*). The relative age relations between the greenstone belts and adjacent banded gneisses are frequently obscured through deformation and can only be determined by precise zircon dating (*22*). The original depositional environment in greenstone belts is still disputed, and models range from entirely intraoceanic (oceanic crust or oceanic plateau) to oceanic ridge–forearc basins, marginal basins, island arcs, active continental margins, continental platforms, and continental rifts (*7*). The time span for early Archean greenstone evolution may vary considerably from belt to belt, although periods of about 200 to 250 Ma (million years) appear reasonable (*23*).

3) The greenstones are intruded by granitoid sheets and plutons, largely of TTG composition but also comprising late, K-rich granites and pegmatites, many of which are associated with gold mineralization (*16*). Geochemical data and zircon ages show that many of these granitoids are contemporaneous with felsic volcanism in the greenstone sequences (*23, 24*). In many areas these plutons are oval-shaped composite batholiths and stand out spectacularly on satellite photographs.

Crustal Growth and Differentiation

Archean crust (exposed and buried) is estimated to make up about 7% of the present continents (Fig. 1) (*1*). Geochronologic data suggest that much of this crust was generated in distinct pulses, sometimes referred to as Crustal Accretion and Differentiation (CAD) events (*25*), but the occurrence of distinct pulses may be an artifact due to insufficient and nonrepresentative dating. Geochemical and isotopic modeling suggests that ~40% of the present continental volume had been generated by about

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3.8 Ga ago (*26*) and had grown to at least 50% by about 2.5 Ga ago (*27*).

Rapid early crustal growth leading to gneiss-granite-greenstone terranes up to several hundred kilometers in extent by the end of the Archean, 2.5 Ga ago, has been ascribed to crust accretion rates up to three times those observed today (*28*). This growth is ascribed to high heat flow in early Earth, resulting in fast plate motion and correspondingly fast plate turnover (*29, 30*). Radioactive decay requires that Archean terrestrial heat flow was two to three times the present rate (*31*), and the decrease in heat flow during the Proterozoic and Phanerozoic would account for a corresponding decrease in the rate of crust formation through time (*32*). Nevertheless, there is still considerable disagreement on the precise mechanism of crustal accretion (*7*).

Although crust older than 3.5 Ga is only known from a few places in the Archean cratons and is volumetrically insignificant in comparison to the amount predicted by Nd isotopic data, there is compelling indirect evidence suggesting that continental lithosphere had already grown to substantial thickness by about 3.5 Ga ago. Even the earliest known mantle-derived rocks up to 3.96 Ga in age contain an isotopic signature indicating that a significant part of the originally chondritic mantle was already depleted in Nd (*33–35*), probably as a result of voluminous crust formation. Detrital zircons from metaquartzite in the northern Yilgarn craton of Western Australia crystallized 4.2 to 4.3 Ga ago (*36*) and so far constitute the only evidence of this early crust at a time when Earth, like the other planets, was still heavily bombarded by meteorites (*37*). Tonalitic gneisses 3.96 Ga old from the Slave Province of Canada (*38*), equally old detrital zircons found in metasediments of the Wyoming Province in Montana (*39*), and tonalitic gneisses 3.8 to 3.9 Ga old in West Greenland and Labrador (*8*) indicate that segments of continental crust of unknown dimensions were in existence by that time.

This view, among others, is supported by significant intracrustal differentiation shown by negative Eu anomalies [Eu is one of the rare earth elements (REE)] in the oldest preserved sediments and in pre-3.5-Ga granitoids. The REE are useful for inferring crustal growth processes (*27*) because they reflect petrogenetic processes during crust formation and differentiation. Rocks formed from so-called "primitive" (that is, mantle-derived) melts normally exhibit a smooth REE pattern when normalized to chondrites, whereas intermediate to acid rocks derived from intracrustal melting typically display a pronounced negative Eu anomaly (that is, the rock contains less Eu than expected for a smooth pattern) because much of the Eu is retained in plagioclase remaining in the source material. Many Archean sedimentary rocks, particularly those derived from erosion of greenstone belts, have no negative Eu anomaly; therefore, it was initially concluded that there was little internal crustal differentiation in the Archean and, by implication, that the crust was thin and relatively primitive (*27*). Recently, however, significant negative Eu anomalies have been discovered in early Archean clastic metasediments (*40, 41*) and in pre-3.5-Ga granitoids (*42*), which imply that these rocks were derived from erosion of already differentiated crust. Because crustal remelting is most likely in thickened crust, a revision of the earlier concept is therefore required and favors the presence of a crust probably as thick as today, at least locally, that differentiated through intracrustal melting early in the evolution of the continental lithosphere. Models relating crustal generation, isostatic balance, freeboard, and heat generation also predict that the earliest continental crust had a thickness close to the present average value of about 30 km (*43*).

The presence of a thick crust is confirmed by the preservation of extensive early Archean high-grade metamorphic assemblages. These rocks record lower crustal conditions, and pressure-temperature data suggest that temperature gradients were only 10° to 14°C per kilometer at that time, similar to those in cratonic continental crust today (*44*). If early Archean mantle temperatures were about 200° to 400°C higher than those today, such low temperature gradients would only be possible in a crust shielded from high-mantle heat flow by a thick, insulating layer of subcrustal lithosphere (*45*) in which heat transport was by conduction rather than by convection. The presence of a thick (150 to 200 km) lithospheric root is also supported by thermobarometry on early Archean diamonds that crystallized at such depths (*46, 47*) and suggests overlying continental material at least several hundred kilometers in horizontal dimensions.

The gross discrepancy between currently known Archean crustal volume (Fig. 1) and the continental mass predicted from isotopic data may mean either that much of this crust is still hidden in the ancient cratons, a view not favored by geochemists (*35*), or that this crust was recycled back into the mantle in such a way that it did not substantially affect the isotopic systematics. The only way this seems possible is that this crust is now stored in the subcrustal lithosphere below the old shields.

Archean Plate Tectonics

The question of whether plate tectonics operated in the Archean has been discussed at length in the literature (*1, 7, 48–50*). Unresolved questions are whether the gneiss-granite-greenstone association developed in intraoceanic island arcs or Andean-type active continental margins and whether the mafic-ultramafic rocks of greenstone belts can be related to spreading oceanic crust. Although most researchers now favor some form of plate interaction, no consensus has so far emerged on whether this was similar to what we observe today (*32*). Proponents of modern-style plate tectonics in the early Archean have largely argued their case on the basis of geochemical criteria; in particular, the apparent similarity of the Archean TTG suite to Phanerozoic subduction-related calc-alkaline rocks (*2, 12, 27, 51*) and also on a comparison of

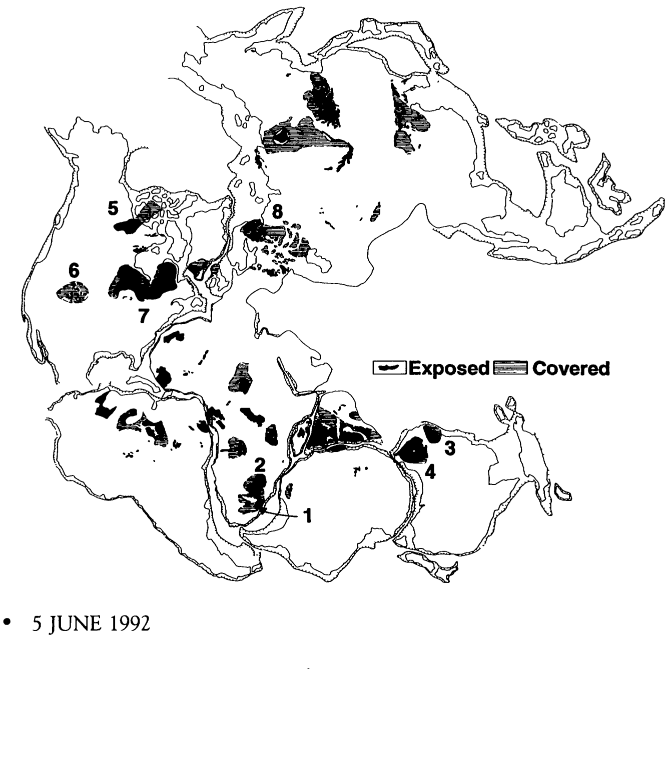

Fig. 1. Distribution of exposed and covered (that is, sub-Phanerozoic) Archean crust as preserved in a Pangaean reconstruction of the continents [adapted from (*1*)]. 1, Eastern Kaapvaal craton including Barberton greenstone belt; 2, Zimbabwe craton (formerly Rhodesian craton); 3, Pilbara craton (also known as Pilbara block); 4, Yilgarn craton (also known as Yilgarn block); 5, Slave craton (also known as Slave Province); 6, Wyoming craton (also known as Wyoming Province); 7, Superior craton (also known as Superior Province); 8, Eastern Baltic shield.

greenstone belt mafic-ultramafic volcanic rocks with modern oceanic basalts (*12*). Some greenstone belt successions have also been interpreted as obducted Archean ophiolites (*19, 52, 53*), but convincing field evidence for Archean layered oceanic crust with sheeted dike complexes is still lacking. Opponents of the above views have noted that andesites are absent or rare in early Archean volcanic suites and that average Archean continental crust is more felsic than modern island arc crust (*54*). Furthermore, TTG rocks have REE patterns unlike their modern counterparts in island arcs (*4, 55*), and sedimentary assemblages resembling those in modern accretionary wedges are rare or absent in early Archean terranes. Nevertheless, the recognition that horizontal tectonics (*17*) rather than deformation through subsidence ("downsagging") of greenstone material between rising TTG diapirs (*1, 16*) characterizes the deformation style in virtually all Archean terranes strongly suggests that crustal shortening, probably resulting from plate convergence, was the dominant process in Archean tectonics (*7, 56*). It is likely, therefore, that some form of plate tectonics operated in early Earth, but the precise mechanism of plate interaction remains speculative.

The issue of Archean plate motion also remains controversial, because most arguments are based on theoretical considerations concerning the thermal evolution of Earth. One view suggests that higher mantle temperatures resulted in higher heat production and that an increased ocean floor spreading rate and a greater ridge length than today was a means of losing more of the heat produced in the Archean mantle (*57*). In contrast, numerical calculations and available paleomagnetic data suggest that plate velocities have not changed much through geologic time (*35, 48, 58–60*). If the latter conclusion is correct, the extraordinarily high crust production rate in the Archean, which was up to three times as high as today's (*28*; Fig. 2), cannot be accounted for solely by lateral magmatic accretion as in modern arcs and active continental margins. It has been suggested that vertical accretion through plume-induced magmatic underplating has been a dominant process of crust formation in early Earth (*7, 48, 61*), a view now supported by geophysical modeling (*62*) as well as petrologic arguments (*63, 64*). For example, mantle-derived melts high in MgO as represented by Archean komatiites have densities higher than most crustal rocks, and such melts would largely be trapped near the base of the crust and form a reservoir from which more differentiated melts, partly contaminated by adjacent continental material, could rise to form basalt or tonalite.

Earliest Crust Formation

High heat production in early Earth implies that a hot early Archean mantle had a chaotic convective regime (*65*), inhibiting the development of a globe-encircling mosaic of thick, rigid plates that interacted at boundaries like those of today. Instead, hot-spot magmatism above numerous plumes may have dominated early Earth (*28, 31, 66*) and may have been the main mechanism of earliest crustal growth (*62*), as has also been suggested for the crust on Venus (*67, 68*). Thermal and petrologic arguments suggest that this earliest crust was far thicker than modern oceanic crust (*32, 35, 69, 70*), although there are also proponents for a thin crust (*14*). Kröner (*7, 48, 61*) and Maalløe (*71*) drew an analogy between the growth of Iceland and the formation of earliest continental crust above a plume, involving massive underplating of mantle-derived magmas and early intracrustal melting to generate granitoid rocks of the TTG suite (Fig. 3).

Intracrustal differentiation in the earliest mafic-ultramafic crust would have been facilitated by the high temperatures associated with plumes because a thick subcrustal "root" had not yet developed. The dense residue remaining from the generation of felsic melts could have either sunk back into the convecting mantle (*72*) or become trapped in the slowly thickening lithosphere. Thickening of the lithosphere probably took place through wholesale underplating of Fe-depleted mantle from which a komatiitic to basaltic melt had been extracted, leaving a buoyant peridotite residue less dense than its protolith below (*73, 74*).

Crustal differentiation above such plumes rapidly led to a significant density

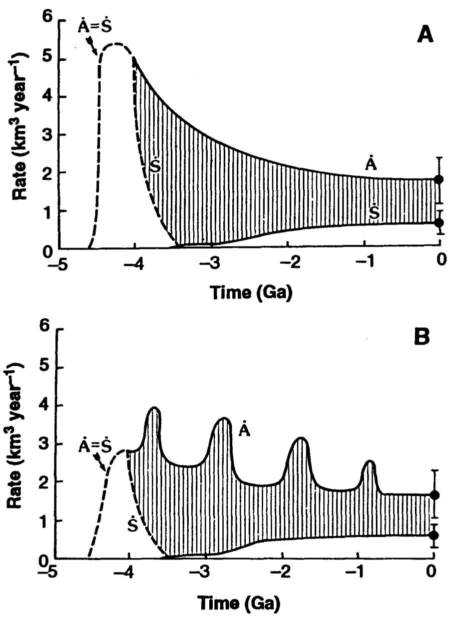

Fig. 2. Hypothetical curves showing crust addition (A) and subtraction (S) rates through Earth history. (**A**) Curve proportional to the decline of heat flow from the mantle. (**B**) Addition rate curve showing episodic periods characterized by high crust addition rates. [Adapted from (*28*)]

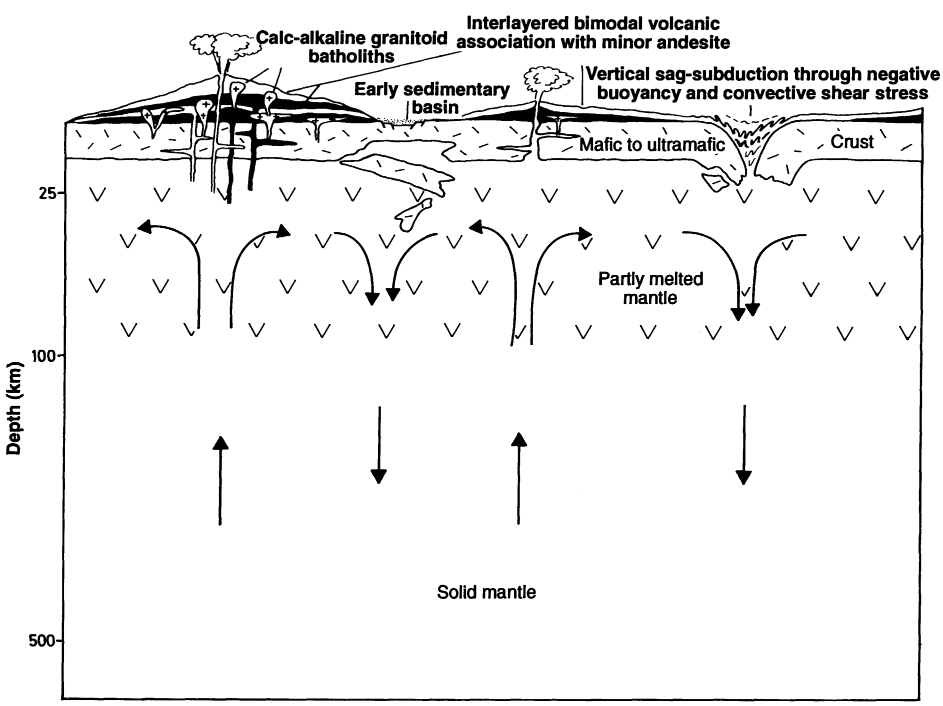

Fig. 3. Schematic and speculative model for the origin and growth of the earliest continental crust, based on an idea in (*10*). [From (*48*)]

difference between the evolving "minicontinents" and the underlying mantle; this difference would have inhibited recycling of this crust back into the mantle through convective drag. In contrast, recycling would have been possible in mafic-ultramafic crust that did not evolve over plumes and therefore could not differentiate.

Galer and Goldstein (35) suggested that early mantle depletion does not require rapid formation of voluminous continental material but that the mantle-balancing, enriched (low Sm/Nd ratio) reservoir was in a globe-encircling thick basaltic crust. Although perhaps plausible for the first 200 million years of Earth history, this model fails to explain why no trace of this enriched basalt is preserved in the early Archean rock record, particularly if it resisted subduction.

Severe impact cratering has been called for to continuously modify whatever crust formed on Earth before about 3.8 Ga ago (*14*), and this process may also have aided recycling of early crust back into the mantle, although the effects of this impacting on crustal recycling remain speculative. It is indeed surprising, as observed by Warren (*14*), that no known Archean terrestrial rock bears the textural sign of impact brecciation such as pseudotachylytes. Such textures will not be destroyed completely by subsequent regional metamorphism and deformation, and even the high-level and low-grade greenstone lithologies bear no sign of impacting. However, that impacts did occur even after 3.8 Ga ago is now documented by quenched spherules recorded in the ~3.4-Ga Barberton belt of South Africa (75).

In summary, we suggest that crust formation and differentiation into what one could call continental material was well under way by 4 Ga, probably in numerous and globe-encircling Iceland-type nuclei up to several 100 km in diameter, and it is unlikely that impacting of large planetesimals inhibited the buoyant granitoid material from aggregating, particularly if the emerging microcontinents developed thick subcrustal roots early in their history.

A major problem with this scenario in a still non–plate tectonic early Earth is how to account for the ubiquitous evidence for severe horizontal shortening as documented in the deformation style of even the oldest crustal rocks (56). A comparison with the tectonics of Venus provides some perspective (68) because this planet's heat loss seems to occur predominantly in hot spots that appear to be linked to zones of rifting and spreading, as we envisage for early Earth (35). Evidence for extensive shortening and crustal thickening, perhaps by imbrication, thrusting, and flake tectonics (76) has been recognized in the broad fold belts in Ishtar Terra (77) that stand up to 4 km above the smooth plain of Lakshmi Plateau. This folding probably resulted from large-scale lateral movement, perhaps as in terrestrial ice floes, and was caused by sideways motion of lithospheric fragments above the mantle plumes. Such "crumpling," if it also occurred above plumes in early Earth, may have further enhanced thickening of the evolving continental domes and facilitated intracrustal melting, leading to the injection of large volumes of sheet-like TTG plutons (9) into the evolving protocontinents.

There is still no satisfactory model for the evolution of early Archean greenstone belts, but present data rule out an entirely oceanic development by island-arc formation on oceanic crust (*12*, *50*). Such a scenario is at variance with the voluminous felsic metavolcanic rocks interbedded with komatiitic and basaltic lavas and contemporaneous TTG magmatism that could hardly have been generated in an intraoceanic environment (78). Furthermore, the ubiquitous presence of xenocrystic zircons in the felsic rocks of ancient greenstone belts argues for the presence of older continental crust below or nearby. Continental margins, marginal basins, or continental rifts may therefore have been more appropriate tectonic settings, and the closure of such basins through collision probably produced the tectonic style now found in these belts.

Archean Plate Motion

Suggestions for Archean plate motion up to one order of magnitude faster than today were based largely on the assumption of high heat flow (29) and a low viscosity in the mantle (79). However, recent numerical work indicates that variable rather than constant viscosity must be expected during mantle convection and, thus, the dependence of convective heat flow on mantle temperature may be much weaker than expected (58). These arguments imply that plate velocities have not changed much through geologic time. Paleomagnetic data provide information on rates of change in paleolatitude and can therefore be used to test these ideas.

Paleomagnetic studies in southern Africa (80–94) have so far produced six well-dated (2σ errors on age less than 50 Ma) paleopoles spanning about 1000 Ma of crustal history from 3450 to 2460 Ma (Table 1). All localities for which reliable paleopoles could be determined are in the Kaapvaal craton except for the Great Dyke, which is in the Zimbabwe craton. From the location of the paleomagnetic pole at a specific time, the paleolatitude of a site can be calculated (Fig. 4A), assuming no local large-scale tilting.

The African data suggest that at ~3450 Ma (during greenstone belt formation in the Barberton area) and at some time in the late Archean (the poorly constrained time of the Modipe Gabbro intrusion in Botswana), the Kaapvaal craton was at polar latitudes; at about 3212, 2875, and 2460 Ma ago it was at intermediate latitudes; and at 3176 and 2687 Ma ago it was at the paleoequator. Thus, there has been significant apparent polar wander during the Archean.

The rate of latitudinal change and, hence, some estimate of the minimum velocity for the Kaapvaal craton can be calculated from the data (Fig. 4B). The average velocity for the period 3500 to 2400 Ma was 17 mm per year, which is similar to Phanerozoic plate velocity estimates (59).

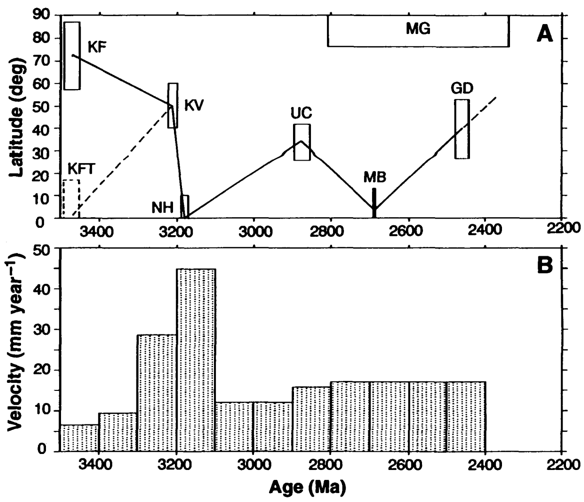

Fig. 4. (**A**) Paleolatitude versus age of paleomagnetic poles from the Kaapvaal-Zimbabwe craton, southern Africa, from 3500 to 2400 Ma. Pole abbreviations are from Table 1. The vertical height of each box represents the A_{95} associated with the pole, the horizontal width represents the 2σ uncertainty in the radiometric age (see also Table 1). The solid line connects the well-constrained poles, and the slope of the line provides an indication of the latitudinal paleovelocity of the Kaapvaal craton (see Fig. 5). The dashed box and line associated with the Komati Formation (tilt-corrected, KFT) assumes that the magnetization was acquired before tilting of the beds, and the pole is corrected for paleohorizontal. (**B**) Paleovelocity versus age for the Kaapvaal-Zimbabwe craton. The velocity is averaged in 100-Ma windows. The average velocity for the time interval 3500 to 2400 Ma is 17 mm per year.

This velocity is a minimum estimate for the following reasons: (i) Changes in paleolongitude are not detected paleomagnetically, so motion parallel to lines of latitude is not considered in the velocity estimate. (ii) With only six poles spanning about 1000 Ma, this data set cannot be used to detect changes in paleolatitude that occur at frequencies less than 100 million to 200 million years. Thus, times when there may have been rapid motion, which was relatively short in duration, may not have been sampled, and the velocity for this interval would be grossly underestimated.

The maximum velocity observed from the data set results from a comparison of paleopoles obtained from the Kaap Valley and Nelshoogte granitoid plutons, both occurring adjacent to each other in the Barberton granite-greenstone terrain. In this study, the younger Nelshoogte paleomagnetic direction is seen as an overprint to the older Kaap Valley pole. Although it is possible that there was tilting of the Kaap Valley pluton before or during Nelshoogte intrusion, structural data indicate that this could not have been more than a few degrees at most (95). The ages of magnetization estimated for these two plutons differ by only 33 ± 10 million years, during which a 50° change in paleolatitude is indicated from the poles. Thus, for this interval, the latitudinal velocity for the Kaapvaal craton is inferred to have been about 168 mm per year. This rate, although higher than the average Phanerozoic plate velocity, is comparable to velocity estimates of the motion of India in the early Tertiary (96).

With the data currently available we conclude that Archean plate velocities were similar to those of later geologic times. These data do not support the hypothesis that the cratons were moving at rates slower than seen today or were fixed at a single latitude (such as the pole) for the entire Archean (97).

In order to test whether plate motion and velocities as seen in the Kaapvaal craton were a global feature of the Archean, reliable paleomagnetic and age data are required from other ancient shields. Most of these areas have not been studied in detail, and paleomagnetic data spanning the full time interval from 3500 to 2500 Ma (for example, North America, Siberia, the Baltics), or sufficient precisely dated paleopoles (Western Australia) are not available. Even though the data set from Australia is sparse, it can be compared with the African results (Fig. 5A). Only three poles from the Pilbara and Yilgarn cratons of Western Australia are dated with the precision necessary to be of use in determining reliable estimates of apparent plate motion (Duffer Formation A, Millindinna Complex, and Widgiemooltha dyke poles, see Table 1). These data suggest that the other poles either have large age uncertainties or are merely dated relative to other rock units in the area.

Comparison of the African and Australian paleolatitude data leads to the following observations. With a few exceptions, the Australian data are consistent with the African paleolatitude versus age curve (Fig. 5A). These exceptions include the presence of poorly dated high-latitude poles that are similar in age to the Modipe Gabbro from Africa. The three well-dated poles from Western Australia correspond temporally to poles in southern Africa. In all three cases the rock types investigated on the two cratons are the same. At about 2400 Ma ago, the Great Dyke was emplaced in the Zimbabwe craton, while the Widgiemooltha and other mafic dikes in Australia were intruded in the Yilgarn craton. At about 2800 Ma ago, two large gabbroic bodies, the Usushwana (Kaapvaal craton) and the Millindinna (Pilbara craton) complexes were emplaced. In both these cases the paleolatitude of the Pilbara craton was identical to that of the Kaapvaal craton at the time of intrusion.

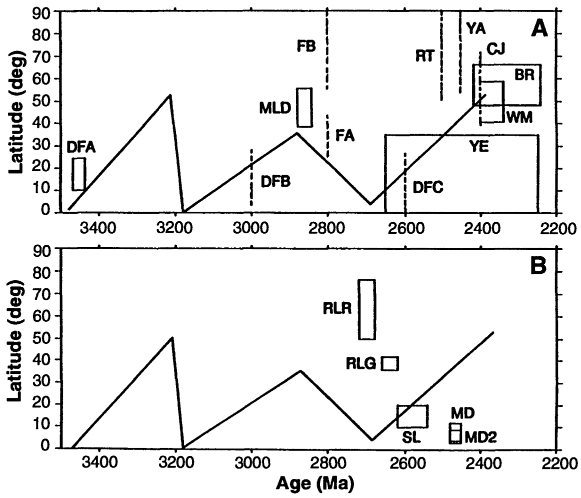

Fig. 5. Paleolatitude versus age of paleomagnetic poles from the Pilbara-Yilgarn craton, Western Australia (**A**), and from the Superior craton, Canada (**B**), from 3500 to 2400 Ma. Boxes are as in Fig. 4A. Pole abbreviations are from Table 1. Dashed vertical lines denote that the age of the pole is not well constrained but is assigned to be approximately the age indicated (see Table 1). The solid line is the paleolatitude curve from the Kaapvaal-Zimbabwe craton from Fig. 4A.

Table 1. Age and paleolatitude from the Kaapvaal-Zimbabwe, Pilbara-Yilgarn, Superior cratons.

Unit	Age (Ma) (±2σ)	Paleolatitude*	Reference
	Kaapvaal-Zimbabwe craton		
Komati Formation (KF)	3470 ± 20	72	(80)
Komati tilt corrected (KFT)	3470 ± 20	2	
Kaap Valley (KV)	3212 ± 24	50	(81, 82)
Nelshoogte (NH)	3179 ± 18	0	(81, 82)
Usushwana (UC)	2875 ± 40	33	(83)
Mbabane (MB)	2687 ± 6	4	(82a)
Great Dyke (GD)	2460 ± 16	39	(84)
Modipe Gabbro (MG)	2575 ± 470	80	(85)
	Pilbara-Yilgarn craton		
Duffer Fm. A (DFA)	3452 ± 16	15	(86)
Duffer Fm. B (DFB)	~3000	9	(86)
Millindinna (MLD)	2860 ± 20	49	(87)
Fortescue A (FA)	~2800	39	(87)
Fortescue B (FB)	~2800	73	(87)
Duffer Fm. C (DFC)	>2600	15	(86)
Ravensthorpe (RT)	2500	71	(88)
YA Yilgarn (YA)	2450	77	(88)
YE Yilgarn (YE)	2450 ± 200	11	(88)
Cajuput (CJ)	~2400	60	(89)
Widgiemooltha (WM)	2370 ± 30	51	(90)
Black Range (BR)	2329 ± 89	59	(89)
	Superior craton		
Red Lake R (RLR)	2700 ± 20	63	(91)
Red Lake G (RLG)	2640 ± 20	38	(91)
Shelley Lake (SL)	2580 ± 40	13	(92)
Matachewan-Hearst (MD)	2450–2480	8	(93)
Matachewan-Hearst (MD2)	2450–2480	7	(94)

*All paleolatitudes are given as positive latitudes (in degrees) and are relative to a point at 30°E, 25°S for the Kaapvaal-Zimbabwe craton; 120°E, 25°S for the Pilbara-Yilgarn craton; and 270°E, 50°N for the Superior craton.

The third case is at about 3450 Ma when the volcanic rocks of the Komati Formation of the Barberton greenstone belt (Kaapvaal craton) and the Duffer Formation of the Pilbara greenstone sequence (Western Australia) were extruded. Both of these formations include basaltic flows with interbedded cherts. In this case, the apparent paleolatitudinal difference is about 50°. At both locations the flows, which were extruded in horizontal layers, are now inclined. Because the flows were rotated during folding, almost certainly after they were magnetized, the preserved magnetic vectors have been rotated as well. When bedding is restored to horizontal, the magnetic vector is rotated to its pretilting configuration, and both the Komati and Duffer formations now show the same paleolatitude of about 0°. For the Komati Formation this tilt correction is shown as a dotted box in Fig. 4A.

The paleomagnetic data, although sparse, are therefore consistent with the idea that the Kaapvaal craton and the Pilbara craton were at similar paleolatitudes for much of the Archean, but the poor age control for the Australian poles makes it difficult to ascertain whether southern Africa and Western Australia acted as a single craton for the whole of the Archean.

New data for well-dated late Archean paleopoles from the Superior craton in Canada (*91*) (Fig. 5B) lead to an average minimum velocity between 2700 and 2440 Ma ago of 25 mm per year, slightly higher than the Kaapvaal velocity. However, as recent studies (*98, 99*) have pointed out, there are disagreements on the ages to assign to various paleopoles and on the direction of motion along the Superior Archean apparent polar wander path.

Recent numerical simulations of continental aggregation and dispersal suggest that supercontinent formation was rare and cyclic through Earth history, and that continental crust was disassembled or fragmented for 70% of the time during the last 3.5 Ga (*100*). A contrasting hypothesis was offered by Piper (*101*), who proposed that there was a supercontinent made up of all continental blocks that existed for most of the Archean. The paleomagnetic results compiled thus far are not able to resolve these two models and permit that there could have been a Pangea-like Archean supercontinent, although probably not with the configuration that Piper (*101*) proposed. The paleopoles for the Superior craton suggest that it was probably not attached to a Kaapvaal-Pilbara craton at about 2700 Ma ago, although more paleomagnetic data are required to confirm this hypothesis. The paleomagnetic data do, however, support the conclusion, also derived from structural studies of Archean rocks (*56*), that plate tectonic processes, similar in style to those since the late Proterozoic, were taking place since at least 3.4 Ga ago.

Conclusions

The broad structural style found in Archean rocks is identical to that observed in younger domains and is compatible with a predominance of horizontal shortening, probably related to plate convergence. However, many of the distinctive features of late Proterozoic and Phanerozoic orogens such as ophiolite-decorated sutures, high-pressure, low-temperature metamorphic assemblages, and fault-bounded terranes with different precollisional histories have so far not been identified even though erosion levels are not lower than in younger crustal domains. Future research on the composition and evolution of Archean terranes should therefore search for microcontinents, exotic terranes, convergence zones, and other hallmarks of the plate tectonic process.

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Human Organ Transplantation: Background and Consequences

Joseph E. Murray

The story of the renal transplant program of the Peter Bent Brigham Hospital (now the Brigham and Women's Hospital) in Boston weaves together three distinct threads: the study of renal disease, the phenomenon of skin grafting in twins, and the development of surgical procedures ultimately leading to the use of chemical immunosuppression. The common leitmotiv is one of a single event or report proving to be decisive. Unanticipated consequences of successful human organ transplantation include the reorganization of clinical and nonclinical disciplines, national and international cooperation in organ preservation and distribution, tissue-typing as a marker for disease, redefinition of death in terms of brain function, better understanding of disease processes, and new health care quandaries that result from the scarcity of organ donors.

Although renal transplantation had been performed sporadically during the first half of this century (*1, 2*), planned programs for human organ transplantation started only in the late 1940s. At that time, clinicians in Paris, London, Edinburgh, and Boston began renal transplantation in nonimmunosuppressed human recipients, in spite of warnings and pessimistic predictions of many scientists and experienced clinicians. Both L. Loeb (*3*) and P. B. Medawar (*4*) claimed that human allotransplantation would never be possible because the roots of individuality were so deep and impenetrable.

Bioscientists had difficulty understanding the determined optimism of clinicians who were willing to evaluate any type of treatment that might help terminally ill uremic patients, most of whom were young and otherwise healthy. Tantalizing reports of functioning human renal transplants had surfaced from time to time; these hints of success were further encouragement. Some researchers, working independently in Paris, had produced temporary function of human renal allografts (*5, 6*), and Lawlor and co-workers in Chicago actually published "success" in a patient (*7*), which was later rescinded.

The first two physicians-in-chief at the Peter Bent Brigham Hospital (the Brigham), H. Christian (1912 to 1939) and S. Weiss (1939 to 1942), were interested in renal disease. When G. W. Thorn became chief in 1943, he and his associate, J. P. O'Hare, shared this interest, especially with regard to the relation of renal disease to hypertension. Although the association of high blood pressure with renal disease had been known for over a century, there was no effective treatment for kidneys damaged by hypertension. After World War II, Thorn invited W. Kolff from the Netherlands to demonstrate a dialysis machine that he had developed during his forced confinement by the Germans (*8*). C. W. Walter helped to improve the design (*1*), and thus the Kolff-Brigham "artificial kidney" was devised. It was first used in patients in 1948 and set the stage for extensive, innovative approaches to both acute reversible renal disease and end-stage kidney failure.

Because renal dialysis was to be only a temporary substitute for renal function, it was logical to seek a more permanent therapy. Chronic dialysis was developed 10 years later in 1958 in Seattle (*9*). Earlier, in the 1940s during a Grand Rounds at the Brigham, Thorn stated that the best way to treat hypertension was to remove both kidneys. The entire audience gasped. The seed for the Brigham renal transplant program had been planted.

Skin Grafting in Twins

This thread in the story involves the biological phenomena of monozygotic (identical) and dizygotic (fraternal) twinning. The monozygotic twin experience starts with the treatment of burns, and the dizygotic twin story begins with freemartin cattle.

In 1932, E. C. Padgett of Kansas City reported the use of skin allografts from family and unrelated donors to cover severely burned patients who had insufficient unburned donor sites for the harvesting of autografts (*10*). Although none of these skin allografts survived permanently, they could be lifesaving by remaining long enough to control infection and fluid loss and thus gaining time for the donor sites to

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