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Consolidation of Continental Crust in Late Archaean–Early Proterozoic Times: A Palaeomagnetic Test

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Abstract

The proposition that continental crust consolidated during late Archaean-early Proterozoic times as spatially distinct groups of cratons is tested using palaeomagnetic results assigned to the interval 2900–2200 Ma. Geological data indicate the existence of two nuclei following widespread cratonisation at $\sim$2700 Ma comprising 'Ur' (stabilisation at $\sim$3000 Ma and shallow water cover at 3000–2800 Ma) including central-southern Africa, western Australia and India, and 'Arctica' which included Laurentia, Fennoscandia and Siberia (stabilisation by $\sim$2500 Ma with supracrustal cover beginning at $\sim$2400 Ma). A third nucleus 'Atlantica', comprising much of western Africa and eastern South America, consolidated later (basement stabilisation at 2200–2000 Ma and $\sim$2000 Ma fluvial-deltaic cover). Using this geometrical premise, palaeomagnetic poles assigned to the interval $\sim$2900–2200 Ma from the two older nuclei are observed to plot along two limbs of an APW swathe. Over 80% of the poles plot on a single segment running through the west of Africa and terminating in a tight group over Australia and India with 'Ur' resided in high latitudes and 'Arctica' in low latitudes. The 2900–2200 Ma mean poles from Africa, Australia, Fennoscandia and Laurentia also accord closely with one another and show that only small relative movements occurred between the cratons during these times. This polar grouping does not result from datasets of similar temporal concentration and is due to low rates of continental movement ($\sim$1 cm/year) that are probably the signature of small-scale tectonic regimes. The primeval continental crust consolidated as two linear belts with a configuration approximating to the geoid form and suggesting an aggregation by whole mantle convection. The relative positions of 'Arctica' and 'Ur' at 2900–2200 Ma were only marginally different from their positions within the putative later supercontinent of Palaeopangaea ($\sim$1100–550 Ma) and imply that the Proterozoic eon was characterised by relatively small intracratonic motions close to limits of palaeomagnetic detection.

Key words: Archaean, Proterozoic, supercontinents, palaeomagnetism, geoid.

Introduction

The mobile nature of crustal tectonics during the late Archaean and early Proterozoic is evident from the close spatial relationship of deformed greenstone belts, granite batholiths, and granite-gneiss terranes and is implied by higher rates of radiogenic heat production during these times. Links between Archaean tectonics and contemporary plate tectonics, and the nature of the Archaean-Proterozoic transition, have been very widely debated. Whilst opinion is still divided, the uniformitarian case has been cogently argued and applied in recent years (Windley, 1997) and has mostly developed from evidence for subduction-related signatures in the Precambrian record. However, there are two important geological difficulties that preclude a rigid application of uniformitarian principles to these times. Firstly, there are the geochemical and isotopic signatures that contrast with Phanerozoic times and the ensialic characteristics of many Proterozoic mobile belts (e.g., Engel et al., 1974; Veizer, 1989; Windley, 1993; Condie, 1997). Secondly, spatially related groups of cratons within the supercontinent of Pangaea have comparable Archaean–early Proterozoic geological histories. Since this could hardly be the case if they had been repeatedly dispersed and rewelded, it seems that their early Precambrian proximities were retained until Mesozoic times. The present paper is mainly concerned with this second observation.

Common histories of cratonic development can be inferred from comparable ages of stabilisation identified from the timing of youngest juvenile mantle additions and from the ages of the oldest laterally-extensive platform sediments deposited onto uplifted cratons. On the basis of this evidence three ancient nuclei have been recognised by Rogers (1996). Three shields comprising five cratons are covered by shallow-water supracrustal successions older than 3000–2800 Ma and include the Kaapvaal of southern Africa and the Pilbara of Western Australia (eg., Cheney, 1996). Close correlations with the Dharwar, Bhandara and Singhbhum cratons of India are apparent although presently unconfirmed by U–Pb geochronology (Aspler and Chiarenzelli, 1998). Collectively these cratons are grouped into a protocontinent informally referred to as 'Ur'. Protolith histories (but no preserved supracrustal covers) suggest a wider link to Zimbabwe (Cheney, 1996; Aspler and Chiarenzelli, 1998) and probably to Archaean nuclei in Madagascar and eastern Antarctica (Rogers, 1996).

Primeval crust of Archaean age underlies much of North America, Greenland, Siberia and Fennoscandia. Rogers (1996) groups the first three into the protocontinent of 'Arctica'. Although isolated relicts as old as 3,850 Ma are present, major protolith formation was concentrated at 2600–2400 Ma. A subsequent common history of Laurentia and Fennoscandia falls broadly into four stages comprising compression and conjugate shear at 2600–2450 Ma, basic igneous emplacement in extensional regimes at 2450–2000 Ma, calc-alkaline magmatism and orogeny at 2000–1800 Ma (Trans-Hudsonian, Svecofennian and also the Akitkan in Siberia), and marginal subduction and arc accretion at 1800–1500 (Park, 1997). 'Arctica' comprises the cratonic assemblage defined as 'Kenorland' by Williams et al. (1991) whilst the two protocontinents grouped as 'Ur' and 'Arctica' are the same as divisions recognised by Aspler and Chiarenzelli (1998) from analysis of supracrustal histories and mineral provenances. These latter authors identify a temporally distinct development of the two protocontinents until glaciogenic deposition at 2400–2200 Ma.

In addition, at least five cratons in western Africa and eastern South America are significantly younger in age and have a widespread ~2000 Ma fluvio-deltaic sedimentary cover which is mostly only marginally younger than the underlying basement (Rogers, 1996). These cratons have been referred to a younger protocontinent of 'Atlantica' (Rogers, 1996) and stratigraphic links between the São Francisco Craton of Brazil and the Kaapvaal Craton in Africa may have commenced by 2490 Ma (Aspler and Chiarenzelli, 1998).

Thus the timings of crustal consolidation and uplift during late Archaean and Palaeoproterozoic times and the provenances of their sedimentary covers identify three protocontinental divisions that are comparable to much younger continental groupings of nuclei within the supercontinent Pangaea. This observation poses an obvious question: did these cratonic nuclei experience relative movements later in Proterozoic times to become fortuitously aggregated together again by early Palaeozoic times, or were relative movements during the Proterozoic eon much less than these same cratons have experienced since the Mesozoic?

The Palaeomagnetic Test

This paper examines whether the proximities of Archaean–early Palaeoproterozoic continental nuclei predicted by these geological correlations are consistent with the palaeomagnetic evidence. Some 25 years ago it was observed that the ~2800–2200 Ma palaeomagnetic poles correlated with a single apparent polar wander (APW) swathe when rotated into a reconstruction not dissimilar from Pangaea (Piper, 1976), an analysis later reinforced by data to the early 1980's (Piper, 1982, 1987). This observation highlighted a clear correspondence of these poles to a unified, and relatively short, APW path with the implication that the early Proterozoic crust formed a quasi-integral body when viewed from the ~1000 km scale resolvable by palaeomagnetism. Since then the age assignments of many of the palaeomagnetic poles have been revised; some are now known to be applicable to later times and the database from most shields has expanded considerably. Results from Africa, Australia and India remain relatively sparse although key new data points have been added and age constraints on many earlier results have been improved. It is therefore important to examine whether this correlation is still apparent.

The time interval to be considered is 2900–2200 Ma, which, with the exception of a few early Archaean results from Australia, southern Africa and India, embraces the oldest part of the palaeomagnetic record. The three minimum requirements for inclusion here are that the pole should have been calculated from magnetisations resolved from demagnetisation study, be interpreted as primary by the authors, and have an age estimated to better than ±200 Ma. Unfortunately it is not possible to apply criteria 2 of Van der Voo (1990) requiring N>24 samples, precision k≥10 and α_{95} <16° and still retain a meaningful number of results to analyse. This is mainly because many of the Fennoscandian results are based on small numbers of samples although demagnetisation and component analysis are apparently satisfactory and age constraints relatively good; elsewhere results are based on a minimum of 13 samples. Compliance with criterion 2 obviously increases confidence in the result and is noted in the summary list of poles in table 1. Criteria 4 (palaeomagnetic field tests that constrain the age of magnetisation) and 6 (presence of reversals) are also included in this table. Although tectonic coherence with the craton is assumed in the selection, structural control (criterion 5) is commonly unable to confirm the direction of magnetisation because the majority of results come from igneous intrusions or metamorphic rocks magnetised during uplift-related cooling. However, since these magnetisations are from extensive basement complexes and of post-tectonic origin, the assumption that they have not undergone major subsequent tilting is usually a reasonable one. The pole data summarised in table 1 may be consulted in the Global Palaeomagnetic Database (available at http://dragon.ngu.no/) with the exception of a few results from the South Indian Shield noted below. There are also some new age assignments applicable to the last revision (2001) referred to in the present text. Within the data framework noted above I have excluded poles with ages inferred from magnetic correlations and used only those with ages defined by stratigraphic or isotopic data. For Laurentia there are multiple studies of the Matachewan and Nipissing igneous episodes; to avoid biassing the short intervals represented by this magmatism, pole sets have been meaned to give the results listed in table 1. Although some poles of supposed Archaean age from Greenland are also included in the database, these are excluded from consideration here because regional isotopic data imply that the magnetisations were acquired during uplift-related cooling following ~1800 Ma tectonothermal activity (see age evidence summarised in Piper, 1985).

An insufficient number of results are linked to high-precision U–Pb or ^{40}Ar–^{39}Ar dates to permit a meaningful test from key poles (Buchan et al., 2000). Hence for a first order assessment the data are divided into three broad age divisions of 2900–2650, 2650–2400, and 2400–2200 Ma by the symbols in figures 1 and 2. The poles from five shields are plotted in present day co-ordinates in figure 1.

Forty Laurentian poles come from the Superior Craton and one each from the Wyoming (13) and Churchill (35) terranes. All but four plot along a single dogleg swathe running between present north and south poles (Fig. 1(E)) and comparable to the former 'Track 6' known in outline

Table 1. Palaeomagnetic poles assigned to the interval 2900–2200 Ma.

No.	Rock unit	Pole position °N	Pole position °E	Age (Ma)	Criteria 2	Criteria 4	Criteria 6	Database Reference
African Cratons								
1	Usushwana Complex (K)	9	347	2875±40	*			377
2	Modipe Gabbro (K)	-33	31	2783-3000	*	*	*	3431
3	Gaberones Granite (K)	35	104	2783±2		*	*	3343
4	Derdepoort Basalt (K)	-40	5	2781±5	*		*	8717
5	Ventersdorp Lavas (K)	-55	355	2699±16				3403
6	Mbabane Pluton (K)	20	106	2687±6	*		*	6216
7	Nyanzian Lavas (T)	-14	330	2680±10	*	*	*	7538
8	Great Dyke Satellites (Z)	23	56	2574±2	*		*	2361
9	Great Dyke (Z)	21	58	2574±2	*			2362
10	Umvimeela Dyke (Z)	21	62	2574±2	*			7553
11	Kisii Group Lavas (T)	7	345	2531±2	*			2505
12	Garauja Gabbro (T)	-27	341	2500±100				1360
13	Kenya Granites (T)	-61	30	2476±50	*		*	8
14	Post-Kavirondian Granite (T)	41	263	2420±60				8122
15	Ongeluk Lavas (K)	-1	101	2222±13	*	*	*	8250
Australia								
1	Millindina complex	-12	161	2925±3	*			308
2	Black Range Dyke	-32	154	2772±2	*	*		1954
3	Cajaput Dyke	-46	146	2772±2	*			1955
4	Mt. Roe Basalt	-52	178	2765±3	*	*	*	311
5	Mt. Jope Basalt	-41	129	2765±3	*		*	309
6	Widgiemooltha Dykes	-8	157	2410±2	*		*	1889
Fennoscandia								
1	Varpaisjarvi Quartz-diorite	64	313	2680±3		*		1314
2	Monchegorsk Pyroxenite	-19	309	2493±7				7423
3	Karelian Dykes	-20	279	2458±18			*	8464
4	Voche Lambina Metamorphic rocks	-15	301	2450±83				7410
5	Burakova (i) Gabbro-diorite	-28	260	2449±1				8522
6	North Finland Intrusions	-18	265	2440±50				6625
7	Burakova Intrusion (ii)	-25	248	2439±22				7425
8	Tolstik Intrusion	-32	305	2437±7			*	8524
9	Main Range Gabbro	-17	305	2436±25				7424
10	Kolvitza Porphyrites	-29	317	2423±3			*	7404

Table 1. *Contd.*

No.	Rock unit	Pole position °N	°E	Age (Ma)	Criteria: 2	4	6	Database Reference
11	North Karelia Intrusions	-41	245	2415±55				7782
12	Hautavaara Gabbro	-41	306	2400±46				7427
13	Akhmalahti Formation	-60	284	2330±38	*			7647
14	Matozero Schists	-42	293	2250±50				7415
15	Matozero Sill	-38	295	2250±50				7418
16	Segozero Sill	-41	311	2250±50				8518
India								
1	Khammano Quartz-Magnetite	6	175	2600±100	*			
2	Quartz-magnetite rocks	47	152	2535±40	*			20
3	Charnockite Belt A1	-8	163	2500-2300	*		*	
4	Charnockite Belt A2	-45	147	2500-2300	*		*	
5	Charnockite Belt A3	-73	75	2500-2300	*		*	
6	Charnockite Belt A4	-18	133	2500-2300	*		*	
7	Charnockite Belt B	16	85	2300-2200	*		*	
8	Madras Charnockites	16	84	2300-2200	*		*	
Laurentia								
1	Hawk Lake Granites	41	170	2888±2	*		*	6394
2	Dundonald Sill	13	201	2800±200	*			2151
3	Monro Formation	-47	283	2800±200	*			2154
4	Kamiskotia Complex	21	182	2800±200	*	*		2152
5	Archaean Rocks Ontario	-68	260	2800±200	*			1614
6	Griffiths Mine Host A	-19	217	2738±5	*	*		5860
7	Griffiths Mine Host B	-30	190	2738±5	*	*		5861
8	Griffiths Mine Ore A	-15	240	2738±5	*	*		5863
9	Griffiths Mine Ore B	28	301	2738±5	*	*		5864
10	Kinojevis Tholeiites	-47	259	2736±33	*			5850
11	Red Lake Greenstone	75	222	2715±15	*			7145
12	Ghost Range Intrusion	7	183	2710±2	*			1676
13	Stillwater Complex (W)	-62	292	2705±4	*	*		3046
14	Poohbah Complex N	-51	240	2700±25	*			1722
15	Sherman Metavolcanics	-37	295	2700±100	*		*	5851
16	Sherman Iron Ores	7	188	2700±100		*	*	5853
17	Steep Rock Iron Ores	19	208	2700±200				5866
18	Otto Stock	69	227	2680±1	*	*	*	2629
19	Moose Mountain A	-46	274	2680±30	*	*	*	5854
20	Moose Mountain B	-8	194	2680±30	*	*	*	5865
21	Adams Mine Ores	-3	199	2675±35	*	*		5857
22	External and Baldhead Granites	27	195	2668±2	*		*	6393
23	Piwitonei Granite	21	226	2665±25	*	*		8392
24	Shawmere Anorthosite	-78	282	2650±100				393
25	Deformation Zone rocks	-77	266	2635±33				7146
26	Chibougamau Greenstone	-61	273	2600±100				2415
27	Shelley Lake Granite	-78	246	2580±20	*		*	1716
28	Burchell Lake Granite	-71	263	2580±30	*		*	1718
29	Archaean Gneiss	-66	223	2550±100	*			6023
30	Poohbah Complex R	-78	270	2550±50	*			1723
31	Ptarmigan Dykes	-42	220	2506±2				8308
32	Adams Mine Metabasalts	56	248	2500±100	*			5858
33	Matachewan Dykes*	-42	241	2453±2	*	*	*	
34	Thessalon Volcanics	47	191	2375±75	*			2078
35	Lorrain Formation 1	-46	268	2350±50	*			8244
36	Lorrain Formation 2	-21	213	2350±50	*			8480
37	Coleman Member	-31	234	2350±50	*			8248
38	Tulemalu Dykes (C)	1	302	2255±15	*		*	1714
39	Maguire Dykes	-9	267	2229±35	*	*	*	8309
40	Nipissing Diabase N1**	-13	267	2219±4	*	*	*	
41	Nipissing Diabase N2***	40	242	2219±4	*	*	*	
42	Semeterre Dykes	-15	284	2216±8	*	*	*	7190

African poles are distinguished by craton of origin as follows: K= Kaapvaal, T = Tanzanian, Z = Zimbabwean. Laurentian poles distinguished as C and W come from the Churchill and Wyoming Terranes respectively; the remainder are from the Superior Craton. *Mean of 14 studies, **Majority group, mean of 10 studies, ***Minority group, mean of 3 studies.

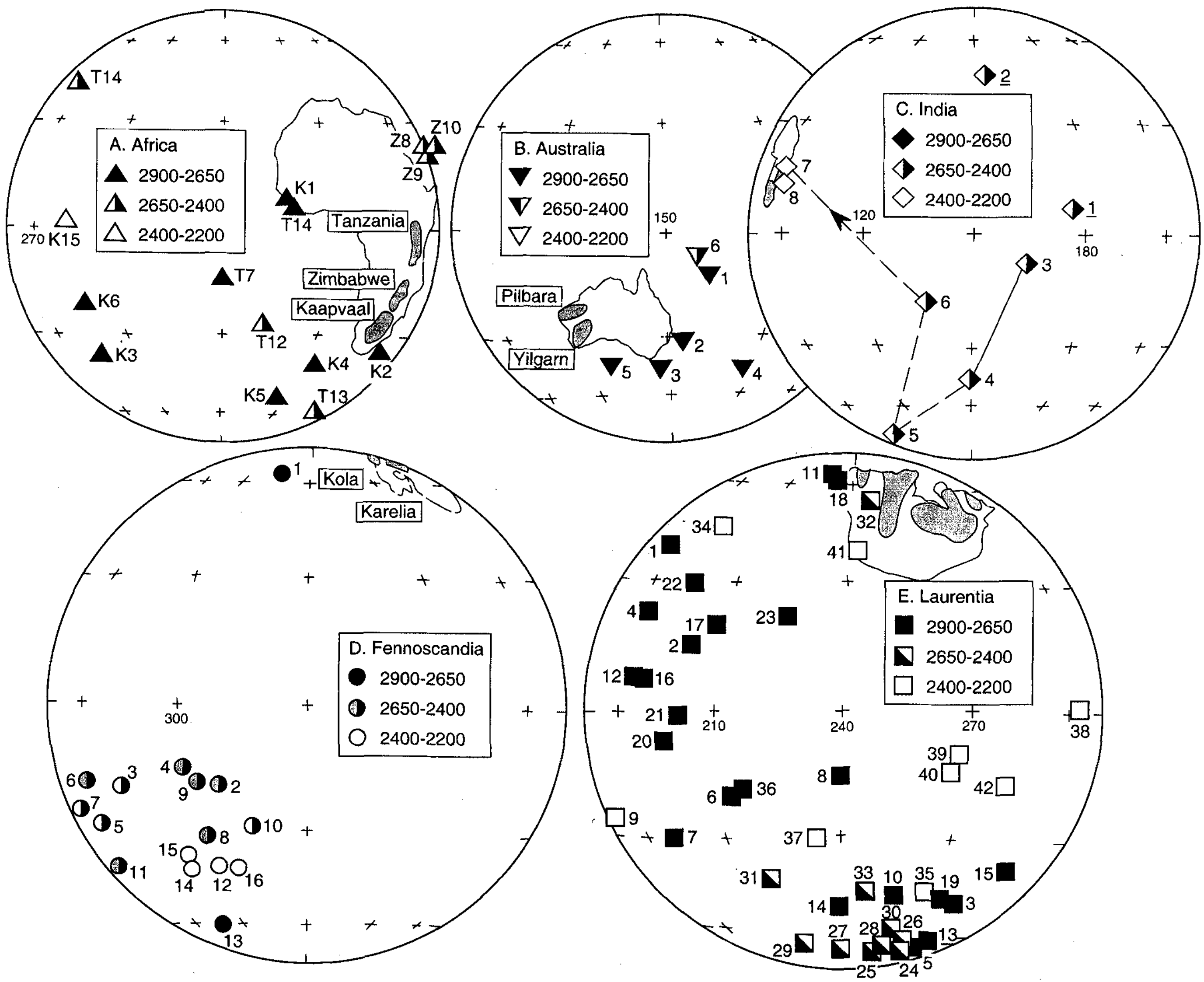

Fig. 1. Summary of selected 2900–2200 Ma palaeomagnetic poles plotted on hemispheric nets (latitude and longitude indicators at 30° intervals) with derivative cratonic nuclei stippled. Poles are plotted in order of age and numbered as in table 1. Fennoscandian results are subdivided into poles from the Kola craton (stippled) and Karelia craton (open).

for more than 25 years (Irving and McGlynn, 1976). The north-easterly segment is poorly recorded and sufficiently discontinuous to identify a polarity ambiguity with poles 11, 18, 32 and 41; reversed equivalents of these could equally well be accepted to plot closer to the tight grouping of poles in the southern hemisphere. There is nevertheless an indication that southeasterly APW is applicable to part of this time interval because all but one of the poles plotting in the north west fall within the oldest age division. Also, extensive studies of Matachewan and related basic magmatism recognise a SE polar motion at ~2450 (e.g., Halls et al., 1994) and the Matachewan pole given in table 1 is a representative mean of a polar distribution plotting between ~230°E, 40°S and 265°E, 50°S. Evidence for more rapid APW motion near the end of the interval under consideration derives from the five poles 38–42 with the youngest age assignments (Table 1 and Fig. 1 E) which relate in part, to the Nipissing igneous episode at ~2200 Ma. Palaeomagnetic studies of this episode recognise two dominant magnetisation directions directed N– and W+ yielding the mean poles 40 and 41 whilst contact and fold tests indicate that both magnetisations are primary (Roy and Lapointe, 1976; Morris 1981; Buchan, 1991). The poles are linked to a range of isotopic ages with a U–Pb baddeleyite date of 2219±4 Ma (Corfu and Andrews, 1986) confirming their assignment to the younger limit of the time interval being considered here. This age determination is probably linked to the N2 magnetisation (pole 41 and Fig. 4) although complex magnetisation structures in the rocks concerned cannot establish this with certainity (Buchan et al., 1989). The general agreement of this N2 pole with subsequent 2200–2000 Ma Laurentian poles is however, a confirmation of a northerly APW movement near the close of the time interval considered here (e.g., Halls et al., 1994 and Buchan et al., 2000).

Expansion in the Fennoscandian shield data set in recent years is based mainly on study of ~2450–2250 Ma igneous episodes in the northern part of the shield and the data are reviewed by Mertanen et al. (1999) and Fedotova et al. (1999). Approximately equal numbers of poles come from the Kola and Karelia cratons separated by the Lapland-Belomorian Belt. The latter includes a 1970–1870 Ma volcanic arc suite interpreted as the site of a collisional suture (Berthelsen and Marker, 1986). On a paleomagnetic scale however, the Fennoscandian poles form a short swathe with no clear distinction between the data sets from the two tectonic provinces (Fig. 1D) and probably representing only a short part of the time interval under consideration. The exception is pole 1 with a much older assigned age.

The short swathe of ~2450–2250 Ma Fennoscandian poles from the Kola and Karelia cratonic nuclei matches the Laurentian swathe in a way which incorporates the SE trend in the 2470–2320 Ma Fennoscandian mean poles identified by Fedotova et al. (1999) and brings divergent pole 1 into the northern limb of the trend (Fig. 2). As with the few Laurentian poles plotting in this area, the polarity of the older pole 1 as plotted in figure 1D is ambiguous. The rotation (–42.5° about a Eulerian pole at 211°E, 60°N) moves the Kola-Karelia nucleus into continuity with the Archaean terranes of Labrador and Greenland (Fig. 3) in a similar proximity to that recognised by other recent analyses (e.g., Mertanen et al., 1999). This conjunction of Fennoscandian and Laurentian Archaean nuclei is also similar to a configuration evident much later in Proterozoic times (Patchett and Bylund, 1977; Piper, 1980; Stearn and Piper, 1984) and supports the interpretation that a single large igneous province embraced the Superior, Kola and Karelian cratons at ~2450 Ma (Heaman, 1997). However, to achieve alignment of the Labradorian and Gothian belts as suggested by geological (Park, 1994) and Mesoproterozoic palaeomagnetic (Buchan et al., 2000; Mertanen et al., 1999) evidence some anticlockwise rotation of Fennoscandia during Palaeo- or early Mesoproterozoic times is implied.

Although only cratonic North America, Greenland and Siberia were assigned to 'Arctica' by Rogers (1996), there is clearly substantial geological (Park 1994, 1997) and

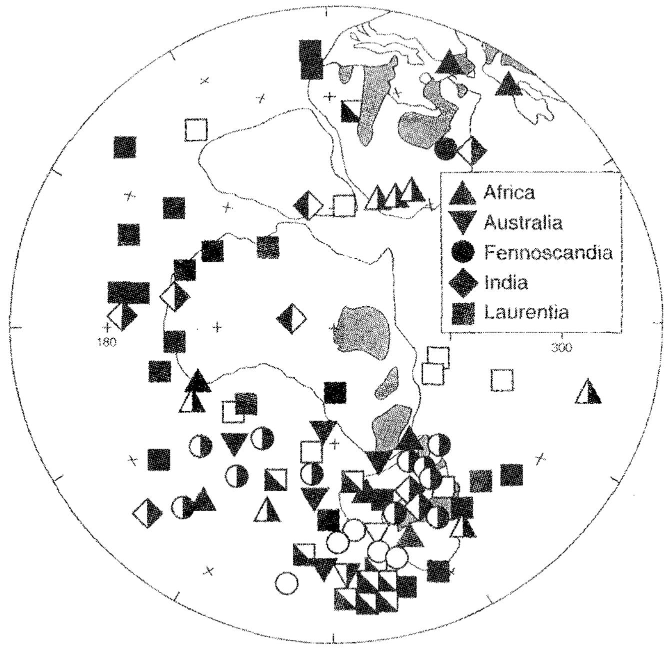

Fig. 2. Palaeomagnetic poles assigned to the interval ~2900–2200 Ma plotted after rotation to bring their distributions into accordance and preserve the proximities indicated by the 'Ur' and 'Arctica' palaeogeographic premises. The resulting continental reconstruction is shown in figure 3. Closed symbols are poles assigned to 2900–2650 Ma, half-open symbols are poles assigned to 2650–2450 Ma and open symbols are poles assigned to 2450–2200 Ma. Rotations (Euler angle positive anticlockwise, Euler pole °E, °N) of cratons to Laurentia are Africa (–146°, 138°, 73.0°), Australia (145.3°, 192.3°, –38.5°), India (–150.5°, 359.0°, –5.5°), Fennoscandia (–42.5°, 211°, 60°).

palaeomagnetic evidence (Buchan et al., 2000; Mertanen et al., 1999 and see Figs. 2 and 4) that a segment of the Fennoscandian Shield was also part of this division by 2500 Ma as included in 'Kenorland' by Williams et al. (1991). Later Mesoproterozoic marginal accretions to Laurentia and Fennoscandia are embraced by the extended protocontinent of 'Nena' (Rogers 1996) and shown in figure 3. The orientation of the Laurentian and Siberian shields during these times is disputed (Condie and Rosen, 1994; Sears and Price, 2000) and cannot be addressed here because palaeomagnetic data from Siberia are currently too poorly reported to test this proposition before Neoproterozoic times. The reassessment of Sears and Price (2000) has noted failure of key tests for a 'Rodinian' position of Siberia north of Laurentia and revived consideration of a location sited to the west of Laurentia. This orientation is adopted in the late Archaean–early Proterozoic reconstruction shown in figure 3 for two reasons. Firstly the poles from Australia all plot close to this nucleus and cannot be matched with the bulk of the Laurentian poles in the conventional Rodinian position west of North America (c.f. Fig. 1B and E). The alternative view developed by matching the polar distributions (Fig. 2) has both Australia and India (mostly in high palaeolatitudes) near opposite parts of the primeval continental crustal body from Laurentia (mostly in low palaeolatitudes). Secondly the adoption of this position from the geological case (although still disputed) produces a broad continuity of continental crust between 'Arctica' and 'Atlantica'.

In the absence of geometrical constraints, poles from the three African cratons (Kaapvaal, Zimbabwe and Tanzania) are retained in present-day orientations in figure 1A. This is a first approximation because terrane accretion was taking place along the western margin of the Kaapvaal Craton at 2900–2650 Ma (de Wit et al., 1992) terminating with consolidation of the Limpopo Belt separating the Kaapvaal and Zimbabwe, probably as these two cratons were welding together (Windley, 1993; Park, 1997). The Zimbabwe and Tanzania cratons are separated by the Pan-African Lufilian-Zambesi Belt. There are broad structural correlations of Mesoproterozoic age across this belt which appear to mitigate against large later dislocations (Shackleton, 1973) although minimum Pan-African motion comprised a sinistral strike-slip (Hanson et al., 1993). In the palaeomagnetic dataset a result from the post-Nyanzian granites (Africa) is excluded on grounds given by Meert et al. (1994) and because this has an apparent age older than the Nyanzian lavas. Results from the Tanzanian Craton now include the pole from the Kisii Group lavas of western Kenya dated 2531$\pm$2 Ma by Pinna et al. (2000). The analysis can also now legitimately include the pole form the Modipe Gabbro (K2). Although error limits on the Rb–Sr determination from this unit exceed $\pm$200 Ma, it is intruded by the Gaberones Granite redated at 2783$\pm$2 Ma (Moore et al., 1993) with a contact test indicating that the magnetisation of the gabbro is older than the granite (Evans, 1967; Evans and McElhinny, 1966). An age assignment of the Modipe pole between 2783 Ma and the oldest K–Ar determination (3000 Ma) is therefore reasonable.

The quasi-integral assumption for these African cratons is supported by the common distribution of the majority of poles plotting on a north to south swathe to the west of Africa (Fig. 1A). If the antipoles of the divergent data (K3, K6, T14 and T15) shown in figure 1A are considered the collective distribution matches the two limbs of the Laurentian-Fennoscandian swathe (Fig. 2) with Africa rotated by –146°, about a Eulerian pole at 138°, 73.0°. There are then five African poles falling on the poorly defined NE segment of the swathe. These include the only data from the Zimbabwe Craton comprising the results from the Great Dyke and related magmatism during a brief magmatic episode at 2574 Ma (Wingate, 2000), and ~2700 Ma Kaapvaal poles K3 and K6. The rotational operation to match the African and Laurentian/ Fennocsandian datasets is identical to that resolved by older assessments of early Precambrian data (Piper, 1980, 1982). It is also the same as the fit of Africa and Laurentia in the much younger supercontinent of Palaeopangaea (Piper, 2000).

Age determinations on the Australian poles have been substantially revised in recent years (Wingate, 1999) but the positions remain to the east of Australia and imply a middle to high latitude location at ~2900–2700 (Fig. 1B). Four of five poles are from rock units within the Pilbara Craton, which is separated from the larger Yilgarn Craton to the south by the Archaean-Palaeoproterozoic Hamersley Basin and Capricorn Orogen. Pole 6 from the Yilgarn does not differ significantly from the Pilbara poles but has a substantially younger assigned age. Collectively these poles match the southern limb of the swathe of data from elsewhere (Fig. 2) with the Australian cratons sited close to southern Africa (Fig. 3). Remarkable geological similarities between the late Archaean geological records of the Kaapvaal and Pilbara cratons have been documented elsewhere (Cheney, 1996; Zegers et al., 1998) but single palaeomagnetic poles do not constrain their relative positions very well (Wingate, 1998). However, because the collective results plot close to Australia they can be matched with the bulk of the data from elsewhere siting the two cratons in close proximity and supporting the geological correlation (Fig. 2). Several configurations are possible within this general constraint and the inverted position of western Australia shown in figure 3 achieves closest proximity between the Pilbara and Kaapvaal. Any radically different attempt to match these results with the poorly defined NE leg of the swathe cannot be made with confidence.

Results from the Indian Shield (Fig. 1C and Table 1) comprise two poles from the Singhbhum Craton in the NE sector of the shield and six poles from the charnockite belt crossing the Dharwar (Karnataka) Craton in central-southern India (Piper et al., 2003). The latter data comprise a sequence of uplift and cooling magnetisations connected in figure 1C that postdates the charnockite event at ~2600 Ma but predates the oldest dyke swarms cutting the shield here (~2370 Ma). Although representative of the uplift related cooling in this belt, they cannot be precisely dated within this range. Minority steep inclination magnetisations from the charnockite belt (poles 7 and 8) are probably younger than poles 3–6 but correspond to similar magnetisations in the oldest dykes (Dawson and Hargraves, 1994). Age limits on the dyke swarms are too poor for inclusion of their poles but the collective data indicate that India lay in high latitudes at

~2300 Ma. The Dharwar and Singhbhum Cratons have comparable Archaean histories and are now separated by a rift zone active during Phanerozoic, and possibly earlier, times (Naqvi et al., 1974). Collectively the poles can be brought into accordance with the two limbs of the data set from elsewhere with the Indian Shield placed adjacent to the Kaapvaal and Pilbara Cratons and compatible with the 'Ur' reconstruction (Figs. 2 and 3). The youngest Indian poles place this continent in high latitudes and correspond specifically with ~2350 Ma Fennoscandian poles at the southern end of the distribution in figure 2. The palaeomagnetic configuration used here is similar to one derived earlier (Piper, 1982) but with India now inverted to match the swathe of charnockite poles to the two limbs of the distribution from elsewhere.

When considered individually, or in small numbers over short time intervals, the palaeomagnetic data from these remote times have a multitude of possible interpretations, many of which have been developed in the literature (e.g., Buchan et al., 2000; Wingate 1998). However, when considered collectively a simple interpretation is possible: the poles assigned to this 700 Ma interval plot on one of two limbs of an APW swathe (Fig. 2), with the great majority defining just one ~110 arc°. In view of the uncertainties surrounding many of the interpretations (which include tilt adjustment in the case of results from the metamorphic complexes and intrusive igneous units), this obvious accordance over a time period appreciably longer than the whole of the Phanerozoic must exclude diverse relative movements between the cratons during these times.

The accordance of the poles is illustrated quantitatively by two calculations. Firstly overall mean poles for each shield derived from the 2900–2200 Ma data sets are listed in table 2 and their rotated positions plotted in figure 4. In spite of the disparity of the age coverage between the different shields, a close correspondence between these mean poles is recognised. Agreement of the Fennoscandian mean with three of the other mean poles could be improved by siting this shield further north around the margin of Greenland. However, the difference is probably real because the result is heavily biassed by ~2450–2350 Ma poles and plots close to the mean 2453 Ma Matachewan pole. High precision U–Pb baddeleyite determinations on basic dyke swarms, including the Matachewan and Suopera, anchor this area of the APW path (Halls et al., 1994; Buchan et al., 2000; Mertanen et al., 1999). Furthermore the mean poles from Laurentia and Africa could be moved further south by considering the antipoles of the data plotting on the NE limb of the APW swathe although the numbers of poles in this area are not large enough to influence the essential conclusion.

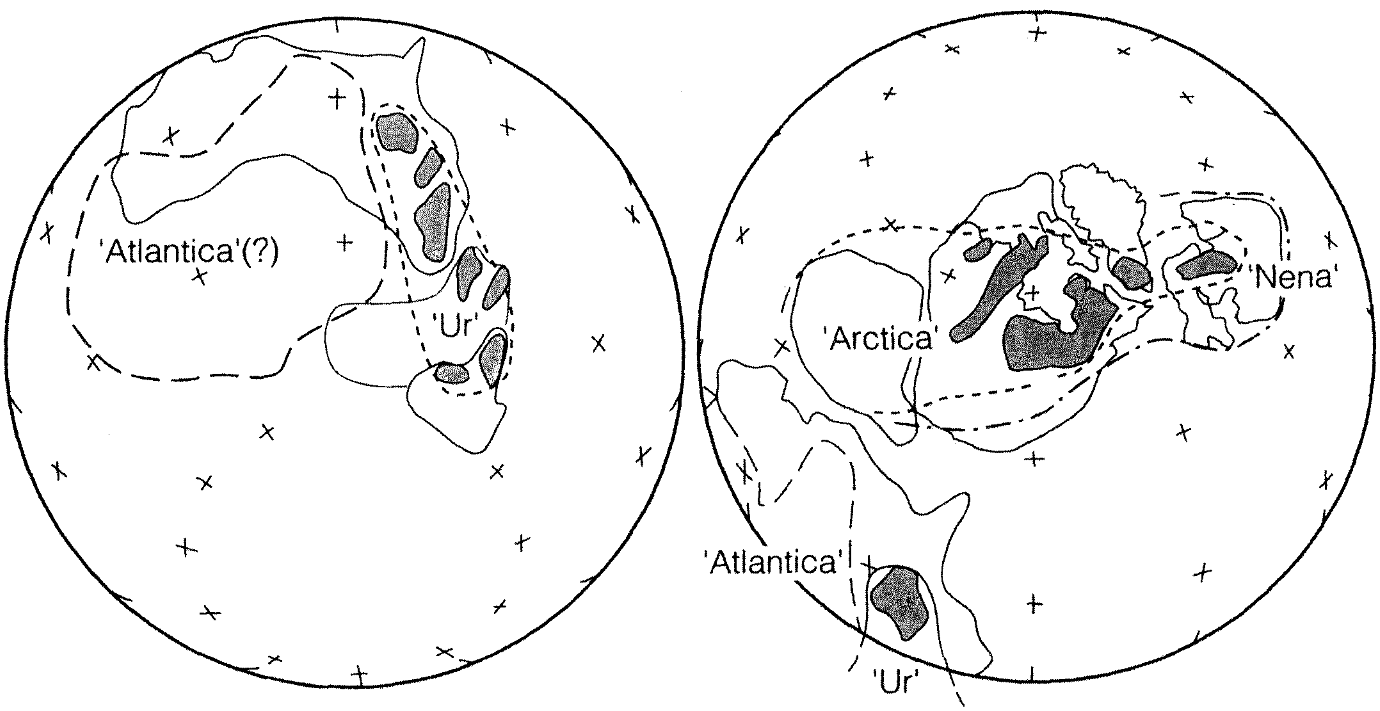

Fig. 3. Palaeogeography of continental crust in late Archaean–early Proterozoic times derived from correlation of palaeomagnetic poles shown in figure 2 and compared with protocontinent distributions indicated by geologic evidence (Rogers, 1996). The location of Siberia is after Sears and Price (2000) and employs a Eulerian rotation of (86.0°, 309.0°, 66.0°).

The averaged pole from the Indian Shield is of little significance (Table 2) because it records the mean of a long distributed swathe of poles (Fig. 1C) but is nevertheless notably close to the mean poles from the other cratons.

Secondly, to gain an indication of APW during this interval a path is calculated from overlapping ±100 Ma data sets through moving 100 Ma time windows (Table 2). Only in the case of Laurentia can this be performed for the whole time interval and the result is a south to north loop indicating slow APW movement prior to 2400 Ma (Fig. 4). Evidence for an increase in the rate of APW after these times rests primarily with the Laurentian data noted earlier but is also suggested by African poles T14 and K15. The APW path calculated in the same way from the combined rotated data from all five shields shows small difference in the pole positions over the interval 2900–2300 Ma with movement characterised by a slow easterly APW motion (Table 2 and Fig. 5).

The tight grouping of mean poles and the small amount of average polar movement imply that APW was low between ~2900 and 2300 Ma with the minority of more dispersed results representing relatively short term departures from a dominant stable mean position. It is an obvious summary of the polar grouping identified in figure 2 and accommodates, but does not explain, the few poles plotting along the NE limb of the swathe. This latter group includes several well-defined data points including results from the Great Dyke and related events in the Zimbabwe Shield (Mushayadebvu et al., 1995; Wingate 2000). As shown in figure 5, a possible accommodation of these results is suggested by the sequence of Indian poles from the charnockite belt, which record an APW loop sometime between 2550 and 2350 Ma.

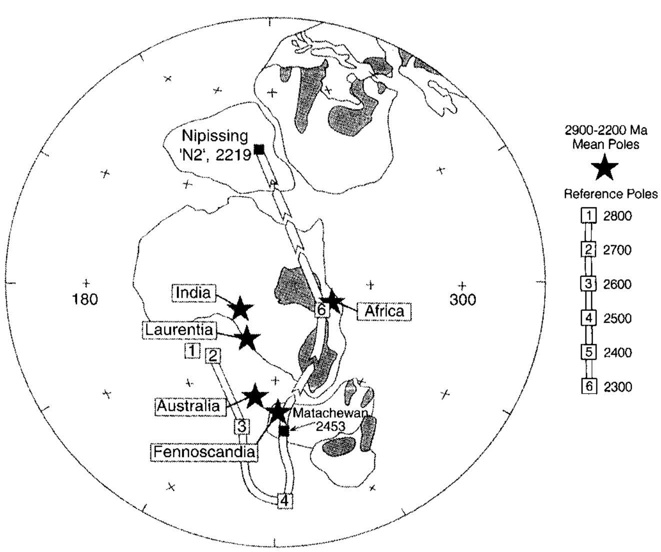

Fig. 4. Mean palaeomagnetic poles from the 2900–2200 Ma data sets from the cratons of southern Africa, Australia, Fennoscandia, India and Laurentia (black stars) rotated according to the reconstruction of figures 2 and 3. Also shown is an APW path for Laurentia calculated from overlapping running means of ±100 Ma data sets.

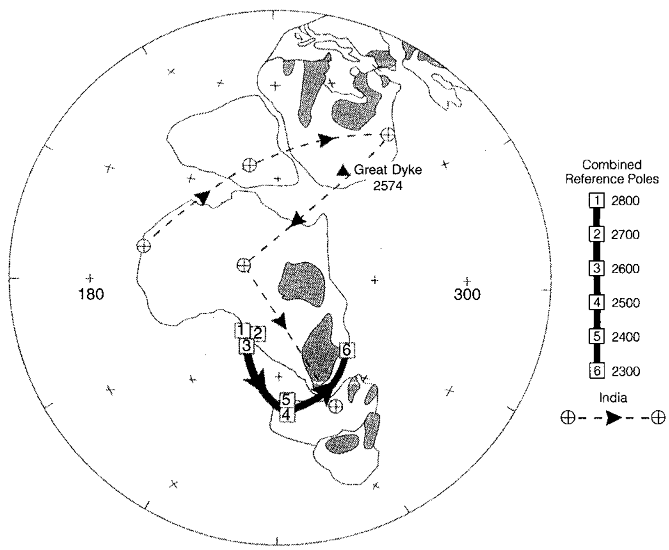

Fig. 5. APW path calculated from the combined palaeomagnetic data from all cratons rotated into the unified reconstruction of figure 3 and calculated from overlapping running means of ±100 Ma data sets.

Rates of Continental Movement During Late Archaean–Early Proterozoic Times

The rates of continental movement are calculated following the method of Gordon et al. (1979) which determines the best minimum estimate from APW movement as a root mean square velocity defined by:

$$V_{RMS} = \sqrt{\int (\omega r)^2 \delta S / \text{area}}$$

Where ω is the angular velocity, r is the position vector on the surface of the Earth and δS is the element of surface area. This function is minimised so that the components of ω satisfy two constraints: firstly $\omega_y = 0$ requires that the Eulerian pole of rotation must lie along the great circle containing the intersection of the palaeoequator and the meridian perpendicular to the great circle connecting two successive mean pole P1 and P2. Secondly ω_x is determined to satisfy the rate of APW as defined by:

$$\omega_x = (P2 - P1)/\ (AGE_2 - AGE_1)$$

The results of two calculations of V_{RMS} are illustrated in figure 6. The first calculates the APWP through 200 Ma time windows incremented in 50 Ma steps whilst the second increments the path in 100 Ma steps. Although absolute values differ somewhat depending on the parameters selected, the general trend is clear (and, as

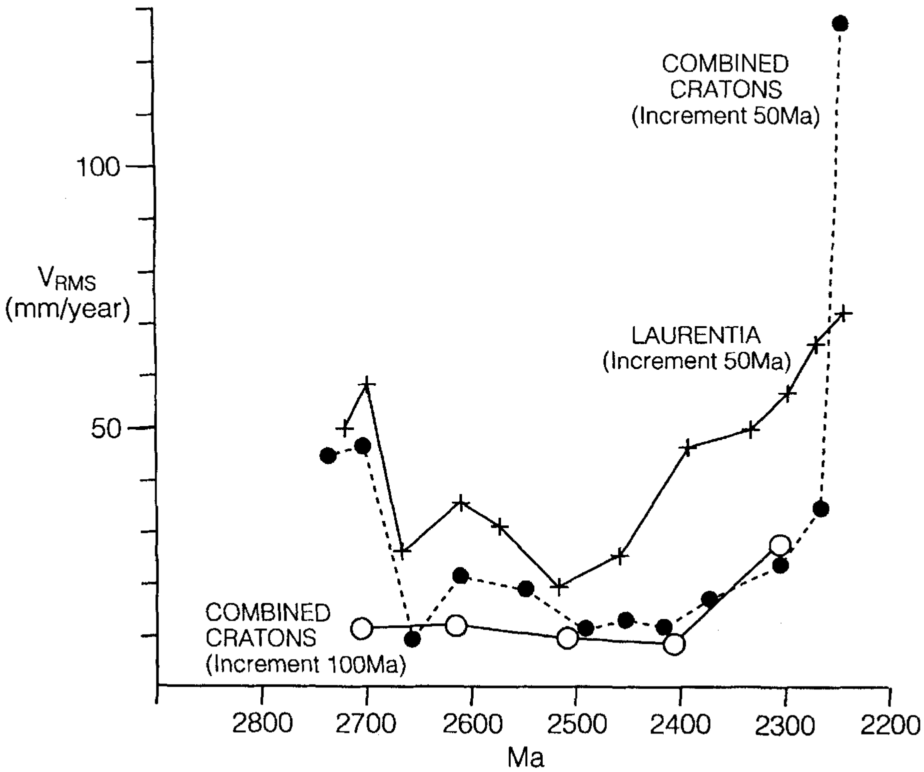

Fig. 6. The root mean square velocity of the continental crust between 2900 and 2200 Ma calculated using the method of Gordon et al. (1979). The velocity determinations are plotted against the mean ages for the combined dataset calculated through ±100 Ma time windows incremented in steps of 50 and 100 Ma and for the Laurentian data set alone incremented in steps of 50 Ma.

Table 2. Mean palaeomagnetic poles for the late Archaean–early Proterozoic shields.

Shield	Time (Ma)	Pole in present coordinates °E	°N	No. of studies	A_{95}	Rotated pole °E	°N
	(a) Overall 2900–2200 Ma Mean Poles						
Laurentia	2900–2200	234	-19	42	18	234	-19
Australia	2900–2200	154	-33	6	19	234	-49
Africa	2900–2200	37	-13	15	33	258	-7
India	2900–2200	132	-8	8	42	227	-7
Fennoscandia*	2900–2200	289	-33	15	12	240	-50
	(b) ±100 Ma Reference Poles						
Laurentia	2800	213	-22	17	31	-	-
	2700	219	-25	25	24	-	-
	2600	225	-44	19	27	-	-
	2500	242	-72	8	19	-	-
	2453**	241	-42	16	5	-	-
	2300***	280	-9	4	21		
	2219****	238	43	3	23	-	-
All Shields	2800	231	-22	23	26		
	2700	233	-21	36	20		
	2600	232	-24	27	23		
	2500	250	-40	27	17		
	2400	247	-40	22	18		
	2300	258	-27	15	26		

Footnote: Laurentian mean in (a) uses polarities of poles 11, 18, 32 and 41 as shown in figure 2(E). African calculation uses antipoles of 6, 3, 14 and 15 as shown in figure 1(A). Fennoscandian calculation excludes pole 1. Mean poles in (b) are derived from a 200 Ma window incremented in intervals of 100 Ma.**Matachewan dyke swarm mean result. *** Calculated from results 38, 39, 40 and 42 in table 1. **** Nipissing N2 mean of 3 results.

shown in figure 6, is present when the Laurentian data are used alone). Velocities of continental movement were low, averaging little more than 1 cm/year during late Archaean and early Proterozoic times. This value compares with rates between 1.5 and 9 cm/year determined from Phanerozoic APW paths using the same method (Piper, 1987). The rate of continental movement certainly increased towards the end of the interval under consideration as the southern part of the continental mass moved towards the pole (Fig. 5) but the case for rates of movement an order of magnitude higher currently depends primarily on the Nipissing Diabase data noted earlier.

Discussion

This analysis has used a broad assessment of 2900–2200 Ma palaeomagnetic data to test the proposition that the oldest crustal nuclei, now preserved within five separated shields, were parts of two distinct assemblages as they were consolidating during late Archaean–early Proterozoic times. The proximities of cratons in southern African, western Australia and India predicted by the 'Ur' proposition are compatible with the palaeomagnetic data although their configurations were evidently different from the later supercontinent of Pangaea. The core of the 'Arctica' proposition is not currently testable due to paucity of data from Siberia, but close links between Laurentia and Fennoscandia are of great antiquity. Between 2900 and 2200 Ma 'Arctica' and 'Ur' appear to have been separated by a distance of the correct size to accommodate the third nucleus of 'Atlantica' (still consolidating during the interval considered here) in its typical Pangaean position (Fig. 3). The apparent offset between the terranes constituting the 'Ur' nucleus in West (southern Africa) and East Gondwana (India and Australia) was observed by Rogers (1996) and can plausibly be linked to Neoproterozoic sinistral strike slip along Pan African belts (Coward and Daly, 1984; Piper, 2000) derived from a primitive reconstruction approximating to figure 3.

The reconstruction derived here from the palaeomagnetic data implies that the primeval crust developed as two linear belts (Fig. 3). The angle between them is comparable to the angle between the two symmetrical limbs of Pangaea, which in turn, matches the form of the hydrostatic geoid (Le Pichon and Huchon, 1984). The same geometrical relationship appears to be reflected in the Neoproterozoic supercontinent Palaeopangaea (Piper, 2000) and is a powerful indication that the large scale mantle processes reflected in the geoid are ultimately responsible for aggregating and constraining continental crust on the surface of the Earth. In the case of the Archaean–early Proterozoic crust the palaeomagnetic poles lie towards one periphery of the continental crust (Fig. 4) and, as noted above, some could be moved further towards the margin by alternative selection of polarities. Hence this primeval body could be superimposed over the geoid negative in a similar way to the Pangaean crust during late Carboniferous to Triassic times as shown by Le Pichon and Huchon (1984).

The analysis can also legitimately conclude that proximities between the cratons remained essentially invariant for the ~700 Ma period considered here. This follows because relative rotations between the cratonic nuclei would distribute the poles derived from lower magnetic inclinations along small circle arcs centred about the sample site, as is observed in neotectonic domains (Macdonald, 1980). Whilst there could be some scope for invoking this on the northern limb of the dogleg in figure 2, it cannot explain why it is not evident on the long and well-defined NW → SE trend.

In view of the many challenges that still face interpretation of palaeomagnetic data during Phanerozoic times, and especially during the early Palaeozoic, it may legitimately be questioned why a simple solution should be apparent over a longer time interval during such a remote era of geologic time. The accordance of mean poles in figure 4 clearly does not result from uniform data coverage over the 2900–2200 Ma interval concerned (Table 1). A hint of the possible reason is apparent in the Laurentian APW path which shows a rapid increase in the rate of APW occurring after ~2300 Ma (Figs. 4 and 6). Hale (1987) proposed that extensive plutonism and crust formation during the late Archaean was linked to enhanced mantle convection which, in turn, permitted cooling of the core and nucleation of the inner core. Additional energy would then become available from the precipitation and sinking of iron and lead to more active core convection. A probable consequence, an increase in the magnitude of the geomagnetic field, is tentatively supported by estimates of palaeofield intensity, which appear to show a large increase near this transition (Hale, 1987). Although there are contrasting interpretations of the link between core changes and mantle convection (Kumuzama et al., 1994; Sleep and Windley, 1994), an increase in the scale of convection, which should be reflected in enhanced APW, is generally regarded as a likely consequence. The transition from Archaean granite-greenstone belt tectonics to Proterozoic mobile/orogenic belt tectonics (Engel et al., 1974; Windley, 1993, 1997; Condie, 1997) is widely considered in this context. Low APW rates seem to have characterised the bulk of the interval 2900–2300 Ma (Table 2) during which 'Arctica' remained in low latitudes and 'Ur' in high latitudes, possibly punctuated by some more rapid short movements (Fig. 5). However, shortly after these times APW rates seem to have increased (Figs. 4 and 6) and onset of the characteristic pattern of APW comprising long tracks separated by hairpin bends is widely recognised subsequently in the Proterozoic (Irving and McGlynn, 1976; Piper, 1982; Fedotova et al, 1999; Buchan et al., 1998). If small-scale tectonic regimes correlate with low APW it might be anticipated that older Archaean palaeomagnetic poles would be little different from the 2900–2200 Ma data set considered here. Four early Archaean palaeomagnetic poles from Africa and India do indeed plot on the NW–SE swathe in figure 2 although data are currently far too few in number to produce a substantial test.

The concept of the Neoproterozoic supercontinent Rodinia (Dalziel, 1991; Hoffman, 1991; Moores, 1991) has done much to stimulate interest in the assembly of continental crust during Precambrian times. Application of the Rodinia reconstruction to the present data set has mixed results. Moving Fennoscandia northwards around the margin of Greenland to make way for South America improves the agreement of mean poles in figure 4 but detracts from the agreement of key ~2400 Ma poles noted above. The majority of African, Australian and Indian poles are shifted away from the Laurentian swathe. This is not unexpected because Rodinia has generally been viewed as an amalgamation of older nuclei during ~1100 Ma orogenesis (Hoffman, 1991). However, an important problem with the conventional reconstruction of Rodinia arises with the Gondwana continents because it separates Australia and India widely from central-southern Africa although these cratons were apparently in proximity during Archaean times (Figs. 2 – 4) and again in Palaeozoic times. Further difficulties with the Rodinia reconstruction cannot be resolved by the 2900–2200 Ma poles but two further problems presented by Rodinia are surmounted by the reconstruction derived here. Firstly the position of Siberia proposed by Sears and Price (2000) is not otherwise required for Australia and Antarctica (Dalziel, 1991; Hoffman, 1991; Moores, 1991). Secondly the belts of mineralisation within Gondwana, which fail to continue into Laurentia as predicted by Rodinia (de Wit et al., 1999), are concentrated within the 'Atlantica' perimeter and are not predicted to continue into adjoining crust.

A final observation to emerge from this analysis is that cratonic configurations of Archaean–early Proterozoic nuclei derived from the correlations of figures 2 are either identical, or differ only by rotations about nearby vertical axes, from their positions within the putative later supercontinent of Palaeopangaea, a Neoproterozoic alternative to Rodinia (Piper, 2000). Since the latter reconstruction was derived from palaeomagnetic data as much as 2000 Ma younger than the data set considered here, a long-term configurative stability of continental crust is evident during Precambrian times. Although palaeomagnetic data for the intervening times await reassessment, it seems that later relative motions between the Archean nuclei (and by implication the crust between 'Ur' and 'Arctica') were small during Proterozoic times. It is often wrongly assumed that the quasi-rigid solution of the Precambrian palaeomagnetic data is in conflict with the well-attested occurrence of Proterozoic ophiolites (Windley, 1993). This is not so. Relative movements detectable by palaeomagnetism have been noted here between Fennoscandia and Laurentia, and between Eastern and Western Gondwana, at times that are as yet, poorly constrained. However, these motions were evidently 'shuffling' in nature. They will have permitted the opening and closing of small ocean basins according to the pattern of Wilson cycles although on a large-scale the continental crust was quasi rigid in palaeomagnetic terms. This is most specifically shown by the comparable reconstruction of the southern African cratons and Laurentia during both the intervals 2900–2200 and 1150 and 540 Ma (Piper, 2000). Models of Precambrian tectonics that fail to accommodate this extraordinary observation can have no long term credibility. The evidence for distinctive geochemical, isotopic and tectonic signatures during the Proterozoic Eon (Garrels and Mackenzie, 1971; Engel et al., 1974; Viezer and Jansen, 1979; Condie, 1997), which has not been widely considered in recent years, therefore merits reappraisal.

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