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Plate tectonics on early Earth? Weighing the paleomagnetic evidence

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ABSTRACT

Paleomagnetism is the only quantitative method available to test for lateral motions by tectonic plates across the surface of ancient Earth. Here, we present several analyses of such motions using strict quality criteria from the global paleomagnetic database of pre–800 Ma rocks. Extensive surface motion of cratons can be documented confidently to older than ca. 2775 Ma, but considering only the most reliable Archean data, we cannot discern differential motion from true polar wander (which can also generate surface motions relative to the geomagnetic reference frame). In order to find evidence for differential motions between pairs of Precambrian cratons, we compared distances between paleomagnetic poles through precisely isochronous intervals for pairs of cratons. The existing database yields several such comparisons with ages ranging from ca. 1110 to ca. 2775 Ma. Only one pair of these ages, 1110–1880 Ma, brackets significantly different apparent polar wander path lengths between the same two cratons and thus demonstrates differential surface motions. If slightly less reliable paleomagnetic results are considered, however, the number of comparisons increases dramatically, and an example is illustrated for which a single additional pole could constrain differential cratonic motion into the earliest Paleoproterozoic and late Neoarchean (in the interval 2445–2680 Ma). In a separate analysis based in part upon moderately reliable paleomagnetic poles, if a specific reconstruction is chosen for Laurentia and Baltica between ca. 1265 and 1750 Ma, then those cratons' rotated apparent polar wander paths show convergence and divergence patterns that accord with regional tectonics and appear to be remarkably similar to predictions from a plate-tectonic conceptual model. Carefully targeted and executed future paleomagnetic studies of the increasingly well-dated Precambrian rock record can imminently extend these tests to ca. 2700 Ma, and with substantially more effort, to perhaps as old as ca. 3500 Ma.

Keywords: plate tectonics, paleomagnetism, Precambrian, Archean, Proterozoic.

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PRINCIPLES AND PRACTICALITIES

Most definitions of plate tectonics include as a major component, if not a complete basis, the lateral motion by large blocks of rigid lithosphere across the surface of the planet. Whereas many of the papers in this volume focus on indirect petrological, geochemical, or isotopic evidence for (or against) early plate tectonics on Earth, this contribution considers the most direct form of evidence for lateral surface motions, consulting the global paleomagnetic record from Archean and Proterozoic cratons. Within a single craton, shifts in paleomagnetic directions from rocks of different ages demonstrate rotations or latitudinal changes of that block relative to the geomagnetic reference frame. When coeval paleomagnetic records are compared between two or more cratons, differential block motion may be distinguished. If paleomagnetic directions vary across a large craton in a manner that is consistent with some assumed geomagnetic field geometry (e.g., a geocentric axial dipole, "GAD"), then internal rigidity of that craton may be inferred. In theory, then, the paleomagnetic method can be used to test the most fundamental kinematic elements of plate tectonics in early Earth history.

In practice, several limitations of the geological record prevent application of paleomagnetic techniques to solve these problems. First, in order to infer geographic movement from observed temporal variations in paleomagnetic direction, one must rely on a specified model for the geomagnetic reference frame. For over half a century, the GAD model has served as the starting point for all paleomagnetic investigations of phenomena occurring at time scales longer than that of typical geomagnetic secular variation ($\sim 10^3$–10^4 yr; Merrill et al., 1996). The GAD model has proved to be consistent, or very nearly so, for time-averaged paleomagnetic directions from rocks formed during the last five million years, of both normal and reverse polarity, within typical errors of ~5° for individual or mean paleomagnetic results (McElhinny et al., 1996; Tauxe, 2005, and references therein).

Determinations of any subsidiary departure from a GAD field in ancient times are limited by the spatial distribution of paleomagnetic data in the context of plate reconstruction models. Pre-Jurassic data cannot be transferred across plate boundaries, for the sake of comparisons and checks on internal consistency, due to the lack of intact oceanic lithosphere from those ages; instead, potential departures from a GAD model must be investigated using single large plates (i.e., supercontinents), presumed episodes of true polar wander (TPW), or latitude-based statistical tests. Inasmuch as supercontinent reconstructions and hypothesized TPW events are based heavily, if not entirely, on paleomagnetic data, the danger of circularity in these tests of field geometry is apparent.

Global statistical tests of pre-Mesozoic nondipole components are limited to zonal field harmonics, and they are further restricted, due to symmetry considerations in the reversing geodynamo, to the geocentric axial octupole component (g3) and its related family of higher harmonics. The magnitude of the octupole component has been proposed to vary between negligible levels and as much as ~20% relative to the dipole during the past 300 m.y. (Torsvik and Van der Voo, 2002). Following the method developed by M.E. Evans (1976) and several subsequent studies, Kent and Smethurst (1998) found a database-wide bias toward shallow inclinations of the entire pre-Mesozoic paleomagnetic record. An obvious candidate for systematic error could be the shallowing of inclinations within sedimentary rocks due to post-depositional compaction, but the bias was also observed within an igneous-only subset of results. The method assumes random sampling of latitudes by the continents as they migrate across the globe, and some recent attention has been paid to this requirement (Meert et al., 2003; M.E. Evans, 2005). As more high-quality data accumulate, there is certainly opportunity to refine this analysis in the future. In the meantime, D. Evans (2006) showed the consistency of paleomagnetic latitudes determined for large evaporite basins from modern times through most of the last two billion years, permitting a GAD field geometry over that interval. Ancient geomagnetic field strength is not well agreed upon, but variations throughout the entire geological record almost always fall within the range of 10%–150% of the present value (Biggin and Thomas, 2003; Macouin et al., 2003; Valet et al., 2005; Tarduno et al., 2006).

In summary, there is diverse evidence from the paleomagnetic record that a modern-like GAD geomagnetic field, whether generally weaker or comparable in strength to that of today, existed throughout the Phanerozoic and Proterozoic Eons, to first approximation. This uniformitarian geodynamo therefore provides a stable frame of reference by which we may measure and compare cratonic motions to assess the likelihood of plate tectonics on early Earth. However, there are more difficult obstacles presented by the geological record, especially for the Archean.

The stability of paleomagnetic remanence in ancient rocks is governed primarily by original ferromagnetic mineralogy and grain size (Butler and Banerjee, 1975) and by subsequent metamorphic conditions. The Curie temperatures of the two most common primary ferromagnetic minerals in crustal rocks, magnetite and hematite, are 580 °C and 675 °C, respectively, roughly within the amphibolite metamorphic facies. However, practical experience from the last half-century of paleomagnetic work has demonstrated that even lower-middle greenschist metamorphism is likely to erase most or all of the earlier magnetic remanence history. This is largely due to the integrated time-temperature history of prolonged tectonothermal activity, which is able to cause superparamagnetic relaxation (resetting) of ferromagnetic mineral grains well below their Curie temperatures (Pullaiah et al., 1975). In a global perspective, subgreenschist stratified Archean rocks are rare. (Unlayered intrusive rocks pose difficulties for paleomagnetic study due to the possibility of unquantifiable tilting during exhumation.) Notable exceptions are the low-grade cover successions of the Kaapvaal and Pilbara cratons, which will be described later, and isolated prehnite-pumpellyite domains within Neoarchean greenstone belts. For a successful paleomagnetic test of relative motions between two Archean crustal blocks, one must not only find the rare example of a subgreenschist stratified unit on the first block, but also the equally rare example of the same age on the second block. Such possibilities become vanishingly small within the preserved geological record on Earth older than ca. 3 Ga. Indeed, as will be shown later, even the relatively rich Proterozoic paleomagnetic record yields few valid tests of relative motions using this technique.

The predictive power of plate kinematics is founded on the internal rigidity of large regions of lithosphere; kinematic data obtained from one side of a plate can then be extrapolated to the remainder of its area. Large sedimentary basins and extensive dike swarms with internally coherent geographical outcrop patterns can demonstrate lithospheric rigidity of cratons. Such features are common in the Proterozoic rock record but dwindle abruptly further back into the Archean record, especially older than 2.7 Ga (e.g., Eriksson and Fedo, 1994; Ernst and Buchan, 2001). Other papers in this volume address whether this change represents secular evolution of tectonic style or merely a degree of preservation of the oldest crust. For our purposes, tectonic coherence of adjacent crustal blocks must be treated as "suspect" in the manner of terranes in accretionary orogens, and the lateral extents of cratons subjected to our paleomagnetic tests steadily decrease with age. For example, rocks of the Abitibi greenstone belt, Canada, can be considered *a priori* as rigidly connected to all of Laurussia during 0.2–0.4 Ga, or to the successively accreting portions of Laurentia at 0.4–1.8 Ga, but only as far as the Superior craton at 1.8–2.0 Ga, only the eastern Superior craton at ca. 2.0–2.65 Ga, and only the Abitibi subprovince itself for the pre–2.65 Ga history (see Hoffman, 1996; Pesonen et al., 2003; Halls and Davis, 2004; Buchan et al., 2007a; Percival, 2007). These areal constraints further limit our ability to apply quantitative paleomagnetic tests to the most ancient geological record.

METHODS

The present analysis of differential motions between Precambrian cratons employs three independent techniques, all of which require pairs of precisely coeval paleomagnetic poles from the same two cratonic blocks, as well as the assumptions of a pure geocentric-axial-dipole magnetic field and a constant planetary radius. The first method relies on an intact amalgamation of cratonic blocks that has persisted to the present time. As long as those blocks have traveled together rigidly, their paleomagnetic apparent polar wander (APW) paths should align perfectly without applying any relative rotations. With increasing age, a complete paleomagnetic data set would document the timing of amalgamation of the individual tectonic elements by way of discordant paleomagnetic poles from ages prior to that assembly. Timing of assembly would be bracketed between the oldest concordant poles and youngest discordant poles when viewed in present coordinates. In the conceptual diagram of Figure 1, the current amalgamation of cratons A + B persists back to the age of their suturing at time t3. This same age should mark the initial convergence of APW paths from each of the blocks A and B in their present positions. A modification of this technique holds true if the now-separated cratons and their APW paths can be restored confidently into a prior period of

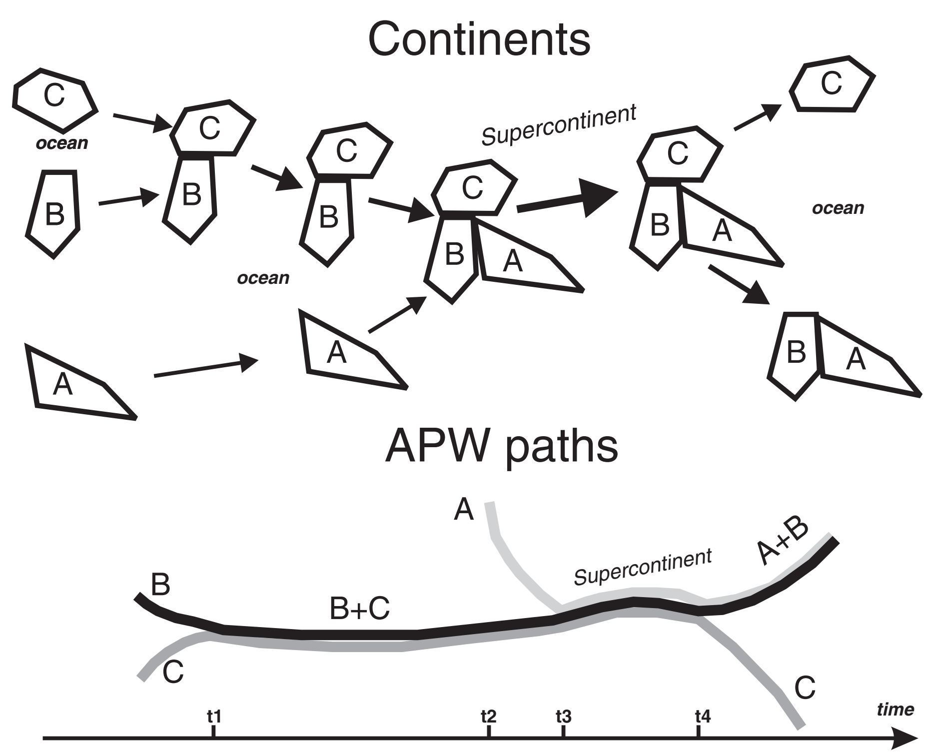

Figure 1. Schematic cartoon of the paleomagnetic record of supercontinental amalgamation and dispersal, involving cratons A, B, and C. Time increases from left to right, with collisions at ages t1 and t3, rifting at age t4, and an arbitrary reference age for discussion in the text at t2. The degree of relative separation is indicated vertically, and this simplified illustration does not discriminate between translational versus local-rotational components of relative motion between cratons. In this idealized example, each apparent polar wander (APW) path is considered to consist of a dense data set of reliable paleomagnetic poles. This figure was inspired by McGlynn and Irving (1975).

assembly such as the Gondwanaland or Pangea configurations of major continental blocks in the early Mesozoic, which have been precisely reconstructed by seafloor-spreading histories (e.g., Besse and Courtillot, 2002). In this case, the APW path of craton C can be rotated to A + B via the same total-reconstruction parameters that reunified the supercontinent, and the age of craton C's initial suturing (t1) is similarly indicated by its converging APW path at that time (Fig. 1).

The remaining technique can be used on cratonic elements for which a specific ancient tectonic association has not yet been established. By this method, great-circle distances between paleomagnetic poles of differing ages are calculated for a given craton, and these are compared with distances between poles from the isochronous interval on another craton. If the angular distances are significantly distinct between blocks, over precisely the same interval of time, then the blocks must have been in relative motion at some time during that interval. Using the schematic diagrams of Figure 1, suppose that cratons A and C, now on separate plates, are constrained by paleomagnetic data at the precise ages of t2 and t4. The difference in total APW path length between those cratons (that of A being significantly longer) proves that they could not have traveled together as a rigid block for that entire interval of time (t2 to t4). In general, this method cannot distinguish between relative translations and vertical-axis rotations; it simply determines whether we can reject the null hypothesis that the two blocks were part of the same kinematic entity. If for those same cratons, paleomagnetic data are available only at times t3 and t4, such a null hypothesis cannot be rejected; however, existence of the supercontinent is not required by these data alone because independently moving plates can generate APW paths of equal length. This test is quantified by the precisions of ages and paleomagnetic pole locations.

To begin our analyses, we considered all ~1500 entries for rocks older than 800 Ma, as listed in the global paleomagnetic database (version 4.6; Pisarevsky, 2005, and available at http://dragon.ngu.no; supplemented by ~20 subsequently published results). We chose this minimum age limit because of near-universal agreement among geoscientists that the hallmarks of modern plate tectonics can be seen at numerous places in the geological record back to at least ca. 800 Ma (Stern, 2005), and also because of the practicality in treating a relatively small proportion of the nearly 10,000 published paleomagnetic results in the global database. Because the implications of our tests are so far-reaching, i.e., a positive result *demands* differential motion between cratons, we chose to limit our discussion to only the data that (we suspect) nearly all paleomagnetists would agree are reliable, providing a definitive basis for our conclusions. To achieve this goal objectively, we queried the global paleomagnetic database with strict quality filters.

Our selection of paleomagnetic poles required satisfying *all* of the following criteria. First, we required precise age of the rock within limits of ±15 m.y. For relative plate translation rates of ~5 cm/yr (typical for continents since breakup of Pangea), these maximal error bounds of 30 m.y. of motion will cause 1500 km of separation, or ~15° of arc at Earth's surface. Typical angular uncertainties on the most reliable paleomagnetic poles are ~5°–10°, and linear addition of two of these uncertainties results in an equivalent amount of error as provided by the age limits. Ages of the most precisely dated paleomagnetic poles have precisions less than a few m.y., afforded by recent advances in U-Pb geochronology. We excluded ages from K-Ar, Rb-Sr, and Pb-Pb on sedimentary rocks, which, in some instances, may have tight precision but dubious accuracy. Second, we required reasonably precise statistics on the mean direction, with radius of 95% confidence cone (A_{95}) less than 20°. Although this is moderately lax, our analysis considers the angular uncertainties quantitatively, and we merely want to distinguish results that likely represent a single directional mode rather than mixed or highly scattered results. Third, we selected only data treated by modern demagnetization techniques including least-squares principal component analysis or, at the very least, vector subtraction. Results from stable-endpoint analysis or "blanket" cleaning were included only when the rocks have been subjected to at least a pilot restudy using more advanced methods. Fourth, at least one positive field test must have been performed to confirm the antiquity of paleomagnetic remanence. Not all field stability tests guarantee primary remanences, and we have included some results in Table 1 that conceivably could be (ancient) remagnetizations. Nevertheless, two-thirds of the selected results in Table 1 have passed tests that imply a primary origin of remanence (baked-contact test, syndepositional fold test, or intraformational conglomerate test), and *all* of the final subset of key poles in Table 2 are among these. Fifth, the rocks must be firmly attached to a craton, with little possibility for local vertical-axis rotation (which could greatly affect pole distances). Also in this general category, plutonic rocks must be assessable for structural tilt. This requirement accepts parallel dike swarms and layered plutonic complexes, but it excludes most granites, gneisses, and deformed mafic lenses within orogenic belts. The discerning reader will have identified elements similar to Van der Voo's (1990) criteria 1–5 in the preceding discussion. We did not require dual polarity of remanence (his criterion 6) or dissimilarity to younger portions of the apparent polar wander path (his criterion 7). Our selection criteria are also similar in spirit to the "key pole" approach of Buchan et al. (2000, 2001).

We selected about fifty results that passed all five criteria, representing only ~3% of the published poles from the pre–800 Ma time interval (Table 1). Our conclusions are therefore on the conservative side for interpretations of differential surface kinematics from paleomagnetic data. It should be noted that despite this low rate for acceptance of results among the entire paleomagnetic database, more than half of the entries on our high-quality list were generated within the past ten years; thus, the rate of high-quality paleomagnetic pole generation has been increasing, in part due to greater recognition among paleomagnetists of the need for adequate statistics and field stability tests, and in part due to the greater precision of isotopic age determinations (principally U-Pb) from the Precambrian geologic record. Also, on an optimistic note, almost 400 of the results that we evaluated were

TABLE 1. PRECISELY DATED PRE–800 Ma PALEOMAGNETIC POLES FROM STRATIFIED ROCKS OR UNDEFORMED DIKE SWARMS IN STABLE CRATONIC BLOCKS

Rock unit (craton)	Age (Ma)	Test*	Pole (°N,°E)	A_{95}°	Pole or age reference
Kandyk suite (Siberia)	1005 ± 4	f	–03, 177	4	Pavlov et al. (2002)
Bangemall sills (Australia)	1070 ± 6	C	34, 095	8	Wingate et al. (2002)
Lake Shore traps (Laurentia)	1087 ± 2	f, C	22, 181	7	Diehl and Haig (1994)
Portage Lake volc. (Laurentia)	ca. 1095	f, G	27, 178	5	Hnat et al. (2006)
Logan sills mean (Laurentia)	**1109 +4/–2**	**C**	**49, 220**	**4**	**Buchan et al. (2000)**
Umkondo mean (Kalahari)	**ca. 1110**	**C**	**64, 039**	**4**	**Gose et al. (2006)**
Abitibi dikes (Laurentia)	1141 ± 2	C	43, 209	14	Ernst and Buchan (1993)
Sudbury dikes (Laurentia)	1235 +7/–3	C	–03, 192	3	Palmer et al. (1977); Stupavsky and Symons (1982)
Post-Jotnian intr. (Baltica)	**ca. 1265**	**C**	**04, 158**	**4**	**Buchan et al. (2000)**
Mackenzie mean (Laurentia)	**1267 ± 2**	**C**	**04, 190**	**5**	**Buchan and Halls (1990)**
McNamara Fm (Laurentia)	1401 ± 6	f	–14, 208	7	Elston et al. (2002)
Purcell lava (Laurentia)	ca. 1440	f	–24, 216	5	Elston et al. (2002)
Snowslip Fm (Laurentia)	1450 ± 14	f	–25, 210	4	Elston et al. (2002)
St. Francois Mtns (Laurentia)	ca. 1476	g, c, F	–13, 219	6	Meert and Stuckey (2002)
Western Channel (Laurentia)	ca. 1590	C	09, 245	7	Irving et al. (1972, 2004); Hamilton and Buchan (2007)
Emmerugga Dol. (N. Australia)	ca. 1645	f	–79, 203	6	Idnurm et al. (1995)
Tatoola Sandstone (N. Australia)	1650 ± 3	f	–61, 187	6	Idnurm et al. (1995)
West Branch volc. (N. Australia)	1709 ± 3	G	–16, 201	11	Idnurm (2000)
Peters Creek volc. (N. Australia)	ca. 1725	g	–26, 221	5	Idnurm (2000)
Cleaver dikes (Laurentia)	1740 +5/–4	c, C	19, 277	6	Irving et al. (2004)
Taihang dikes (N. China)	1769 ± 3	f, C	36, 247	3	Halls et al. (2000)
Post-Waterberg (Kalahari)	**ca. 1875**	**C**	**09, 015**	**17**	**Hanson et al. (2004); de Kock (2007)**†
Molson dikes B (Superior)	**ca. 1880**	**C**	**27, 219**	**4**	**Halls and Heaman (2000)**
Minto dikes (E. Superior)	1998 ± 2	C	38, 174	10	Buchan et al. (1998)
Bushveld mean (Kaapvaal)	ca. 2050	f	12, 027	4	Evans et al. (2002)
Waterberg sequence I (Kaapvaal)	**2054 ± 4**	**c, F, G**	**37, 051**	**11**	**de Kock et al. (2006)**
Kuetsyarvi lavas (Karelia)	**2058 ± 6**	**G**	**23, 298**	**7**	**Torsvik and Meert (1995); Melezhik et al. (2007)**
Cauchon dikes (W. Superior)	2091 ± 2	C	53, 180	9	Halls and Heaman (2000)
Marathon R pol. (W. Superior)	ca. 2105	C	54, 180	7	Buchan et al. (1996); Hamilton et al. (2002); Halls et al. (2005)
Marathon N pol. (W. Superior)	2126 ± 1	C	45, 199	7	Buchan et al. (1996); Halls et al. (2005)
Biscotasing dikes (E. Superior)	2167 ± 2	C	28, 223	11	Buchan et al. (1993)
Nipissing sills N1 (E. Superior)	**2217 ± 4**	**C**	**–17, 272**	**10**	**Buchan et al. (2000)**
Ongeluk lava (Kaapvaal)	**2222 ± 13**	**G**	**–01, 101**	**5**	**Evans et al. (1997)**
Dharwar dikes (Dharwar)	2366 ± 1	C	16, 057	6	Halls et al. (2007)
Widgiemooltha (Yilgarn)	ca. 2415	c	08, 337	8	Evans (1968); Smirnov and Evans (2006)
Karelian D comp (Karelia)	**ca. 2445**	**C**	**–12, 244**	**15**	**Mertanen et al. (2006)**
Matachewan N pol. (E. Superior)	**2446 ± 3**	**C**	**–50, 244**	**5**	**Halls and Davis (2004)**
Matachewan R pol. (W. Superior)	2459 ± 5	C	–51, 257	9	Halls and Davis (2004); Halls et al. (2005)
Great Dike mean (Zimbabwe)	2575 ± 1	c	21, 058	6	Jones et al. (1976); Wilson et al. (1987); Mushayandebvu et al. (1994); Oberthür et al. (2002)
Nyanzian lava (Tanzania)	**ca. 2680**	**g, f**	**–14, 330**	**6**	**Meert et al. (1994)**
Otto Stock (E. Superior, Abitibi)	**2680 ± 1**	**c**	**–69, 047**	**5**	**Pullaiah and Irving (1975); Buchan et al. (1990)**
Stillwater complex (Wyoming)	2705 ± 4	c	–67, 292	17	Xu et al. (1997)
Fortescue Package 1 (Pilbara)	**2772 ± 2**	**f, g**	**–41, 160**	**4**	**Schmidt and Embleton (1985); Strik et al. (2003)**
Derdepoort lavas (Kaapvaal)	**ca. 2782**	**G**	**–40, 005**	**18**	**Wingate (1998)**
Duffer M component (Pilbara)	3467 ± 5	f	44, 086	7	McElhinny and Senanayake (1980); Van Kranendonk et al. (2002)

*Field stability test abbreviations: f—fold test, F—folding penecontemporaneous with rock formation, c—inverse contact test, C—baked-contact test, g—conglomerate test, G—intraformational conglomerate test. Note: capitalized symbols indicate primary magnetization, whereas lowercase symbols indicate merely ancient remanence relative to the geological feature of the test.

†de Kock (2007) has shown that the ca. 1875 Ma post-Waterberg sills are distinctly older than the Sibasa lavas, so only the mean from the sills, rather than the combined pole (Hanson et al., 2004), is listed here.

TABLE 2. ANGULAR DISTANCES OF PRECISELY COEVAL PAIRS OF PALEOMAGNETIC POLES BETWEEN TWO CRATONS (PRE–800 Ma)

Ages (Ma)	Craton 1	Pole dist. (°)	Craton 2	Pole dist. (°)
Key poles only				
1110–1880	**Superior**	**22 ± 8**	**Kalahari**	**58 ± 21**
1110–2220	Superior	81 ± 14	Kaapvaal	79 ± 9
1265–2445	Superior	71 ± 10	Karelia	87 ± 19
1880–2220	Superior	68 ± 14	Kaapvaal	86 ± 22
Adding a single non-key pole (2680 ± 3 Ma Varpaisjärvi basement, Karelia; –64°N, 133°E, A_{95} = 8°; Neuvonen et al., 1981)				
1265–2680	Superior	70 ± 10	Karelia	71 ± 12
2445–2680	**Superior**	**60 ± 10**	**Karelia**	**88 ± 23**
Adding another non-key pole (2686 ± 28 Ma Allanridge lavas, Kaapvaal; –68°N, 356°E, A_{95} = 6°; de Kock, 2007)				
1110–2680	**Superior**	**20 ± 9**	**Kaapvaal**	**45 ± 10**
1880–2680	**Superior**	**42 ± 9**	**Kaapvaal**	**78 ± 23**
2220–2680	Superior	88 ± 15	Kaapvaal	86 ± 11

Note: Boldface type indicates significantly distinguishable values between the two cratons, requiring their differential motion at some time within the given age interval (according to the inherent assumptions of our methods elaborated in the text).

lacking in only one of the first four reliability criteria, including ~40 in which the remanences have been demonstrated to be primary but merely lack adequate statistics or modern demagnetization techniques. Therefore, even if the global paleomagnetic community were merely to revisit those same rocks to bolster the existing published results, we would anticipate a near-doubling of highly reliable data that could be achieved within a few years.

Among the presently available subset listed in Table 1, we next identify groups of poles from different cratons that are coeval within ten million years of rock age (and presumably primary magnetic remanence, as indicated by the field stability tests). Wherever two cratons share coeval poles from more than one age, those poles are listed in Table 2. This is the subset of results that may be considered by the tests for relative motion as described previously.

Paleomagnetic data from Archean rocks are particularly relevant to a discussion on the origins of plate tectonics on Earth, and we note that only a few Archean results (seven, to be precise) pass the stringent reliability criteria we have set forth. Some of the unreliable or unusable results deserve mention here. A substantial portion of the Archean data are determined from unstratified plutonic rocks (some recent examples include the Kaap Valley pluton, Layer et al., 1996; Mbabane pluton, Layer et al., 1989; and Pikwitonei granulite, Zhai et al., 1994) and so are not considered in this analysis. Although figuring prominently in discussions on possible connections between the Kaapvaal and Pilbara cratons (Wingate, 1998; Zegers et al., 1998), poles from the Millindinna and Usushwana complexes (Schmidt and Embleton, 1985; Layer et al., 1988) are excluded because of poor age constraints and lack of field stability tests on the ages of remanence. The Superior craton has yielded two groups of poles with ages near 2710 Ma, with the most reliable examples from each being the Ghost Range complex (Geissman et al., 1982) and the Red Lake granites (Costanzo-Alvarez and Dunlop, 1993; several distinct plutons showed the same remanence direction, thus decreasing the likelihood of regional tilt biases). The two groups of poles are widely separated, however, and each contains members from various Superior craton subprovinces. No preference is made in this analysis regarding these two groups, among which no single paleomagnetic pole satisfies all five reliability criteria. Reliable data are available from the Tanzanian craton at ca. 2680 Ma (Meert et al., 1994), precisely the same age as results from Superior (Table 1), but, unfortunately, no other age comparisons on reliable paleomagnetic poles can be made between those two cratons. Finally, despite a series of well-dated and moderately to highly reliable poles from the Pilbara craton spanning most of geon 27 (Strik et al., 2003), only a single result of comparable age and reliability is to be found anywhere else in the world, from the Kaapvaal craton (Wingate, 1998), which lacks the additional highly reliable poles needed to test for Archean differential motion of those two cratons. Although two of the eight precise age pairs (highlighted in boldface on Table 1) are Archean, those two most ancient pairs are solitary with regard to the two cratons they represent, and they will be unusable for these tests until they are joined by complementary data.

The distribution of our accepted data is skewed toward a few well-studied cratons: Superior (merged with other cratons in Laurentia at ca. 1750 Ma), Karelia (the core of Fennoscandia and merged with the other components of Baltica at ca. 1800–1700 Ma), and Kaapvaal (merged with Zimbabwe to form Kalahari at least by ca. 2000 Ma, if not earlier). Our conclusions, therefore, might not be globally representative. However, the substantial changes in paleolatitude by these few blocks, including some cases of demonstrably differential motion (see following), indicate that the motions we describe are representative of broad regions across the globe.

RESULTS

Perhaps surprisingly, from the entire database of more than 1500 entries, we obtained only ten paleomagnetic poles from five distinct ages that may be used in our analysis (Table 2). It is important to remember that these ten data points are not the only "key" poles from the pre–800 Ma paleomagnetic database, which number approximately fifty by our evaluation. The present data set is limited primarily by the requirement of two pairs of precisely coeval poles from the same two cratons. Ironically, there are several instances of a well-defined APW path segment for one craton that disappears at the same age that another craton's APW path comes into focus. For example, the Superior/Laurentian craton APW path is reasonably well-constrained between 2200 and 1750 Ma, but is unknown for immediately younger segments, when for instance the North Australian cratonic APW path becomes well-defined between 1725 and 1640 Ma. As the North Australian path dwindles in reliability with younger ages, the Laurentian path regains precision at 1590 Ma and younger intervals. Just a few well-targeted studies could fill these gaps.

From the five precise pairings of "key" poles listed in Table 1, four age comparisons emerge, all involving the Superior (or Laurentian) craton: Kalahari at 1110–1880 Ma, Kaapvaal at 1110–2220 Ma and 1880–2220 Ma, and Karelia at 1267–2445 Ma (Table 2). These poles and cratons are plotted in present coordinates (Fig. 2) for the sake of clarity, because subsequent diagrams show them rotated into other reference frames. The four key pole pairs are illustrated in Figure 3, where all elements have been reconstructed to the paleogeographic grid at the younger of the two ages. The only demonstrably differential motion among these four examples involves Superior and Kalahari at 1110–1880 Ma (Fig. 3A), as seen by the lack of overlap between the ca. 1880 Ma poles when rotated into the common 1110 Ma paleogeographic reference frame. In this instance, the cratonic reconstruction shown in Figure 3A is tectonically meaningless, but the cratons are illustrated nonetheless to assist understanding of the concepts. This analysis does not specify when during the 1110–1880 Ma interval the differential motion occurred, so the conservative interpretation would constrain the antiquity of differential cratonic motion, by this method, merely to somewhat older than 1110 Ma.

In the remaining three tests, the lengths of APW are indistinguishable. At face value, this would appear to permit the hypothesis that the two cratons traveled together throughout each entire interval of time, whether as part of a supercontinent, or experiencing true polar wander, or both. However, such a model is untenable for Superior-Kaapvaal between 1110 and 2220 Ma because it has already been demonstrated (Fig. 3A) that those cratons were in relative motion between 1110 and 1880 Ma. The common APW path length in Figure 3B, therefore, is fortuitous. Superior and Kaapvaal could have been members of a supercontinent between 1880 and 2220 Ma (Fig. 3C), but if so, then about four similarly sized cratons will need to be found to fill the ~60° gap between them. Superior and Karelia reconstruct directly adjacent to each other for the interval 1265–2445 Ma (Fig. 3D), but this direct juxtaposition over a billion years is not permitted by the known assembly of eastern Laurentia, occupying the same space as Karelia, during that interval.

Throughout the 1110–1880 Ma age interval, in which differential cratonic motion has now been established, most of the world's cratons can be considered as internally rigid blocks, as evidenced by large, geometrically coherent dike swarms and epicratonic or platformal sedimentary basins (e.g., Ernst and Buchan, 2001; Zhao et al., 2004). By demonstrating differential lateral motion between internally rigid lithospheric blocks older than 1110 Ma, we can quantitatively refute the hypothesis of Stern (2005) that plate tectonics began at ca. 800 Ma. The present comparisons cannot refute Stern's (this volume) modified hypothesis of stagnant-lid tectonics between 1800 and 800 Ma, because it is possible that all of the demonstrated differential motion (shown here to be within 1110–1880 Ma) could have occurred entirely during the 1800–2000 Ma "proto–plate tectonic" regime proposed in Stern's revised model.

To demonstrate the rapid expansion of this kind of analysis allowed by even a modestly growing paleomagnetic database, we will consider two "non-key" poles that, if considered to be reliable or verified as such in the near future, could push back the global record of relative cratonic motion to the interval 2450–2680 Ma, and prove additional between-craton motions in Proterozoic time. The first result is that from the 2680 ± 3 Ma Varpaisjarvi basement complex in east-central Finland; a positive baked-contact test with younger than ca. 2450 Ma intrusions demonstrates a possibly primary remanence (Neuvonen et al., 1981), but the lack of paleohorizontal control requires omission of this result due to our strict quality criteria. However, if we were to include this single result, then comparison with the venerable yet high-quality and precisely coeval Otto Stock lamprophyre dikes pole from 2680 ± 1 Ma in Superior (Pullaiah and Irving, 1975) allows two additional estimates of APW path length: from 1265 Ma and from 2445 Ma (Table 2; Fig. 4). The first test is negative; great-circle distances between 1265 Ma and 2680 Ma poles are nearly identical between the two cratons (Fig. 4A). The second test, between 2445 Ma and 2680 Ma, demonstrates significantly different APW path lengths, even with the poles stretched to their uncertainty limits (Fig. 4B). This result would appear to indicate that Superior and Karelia could not have lain on the same plate through the Neoarchean to earliest Paleoproterozoic interval.

The second example is from the Allanridge lavas on the Kaapvaal craton in South Africa. These rocks stratigraphically separate the Ventersdorp Group, with ages as young as ca. 2710 Ma (Armstrong et al., 1991), from the overlying Transvaal Supergroup, in which protobasins have ages as old as ca. 2665 Ma (Barton et al., 1995). The lavas have been studied paleomagnetically by Strik et al. (2007) and de Kock (2007), the latter of whom documented a positive intraformational conglomerate test and computed a new, combined mean paleomagnetic pole. The only criterion lacking in this combined result is a precise age determination, but the best estimate of ca. 2686 Ma is

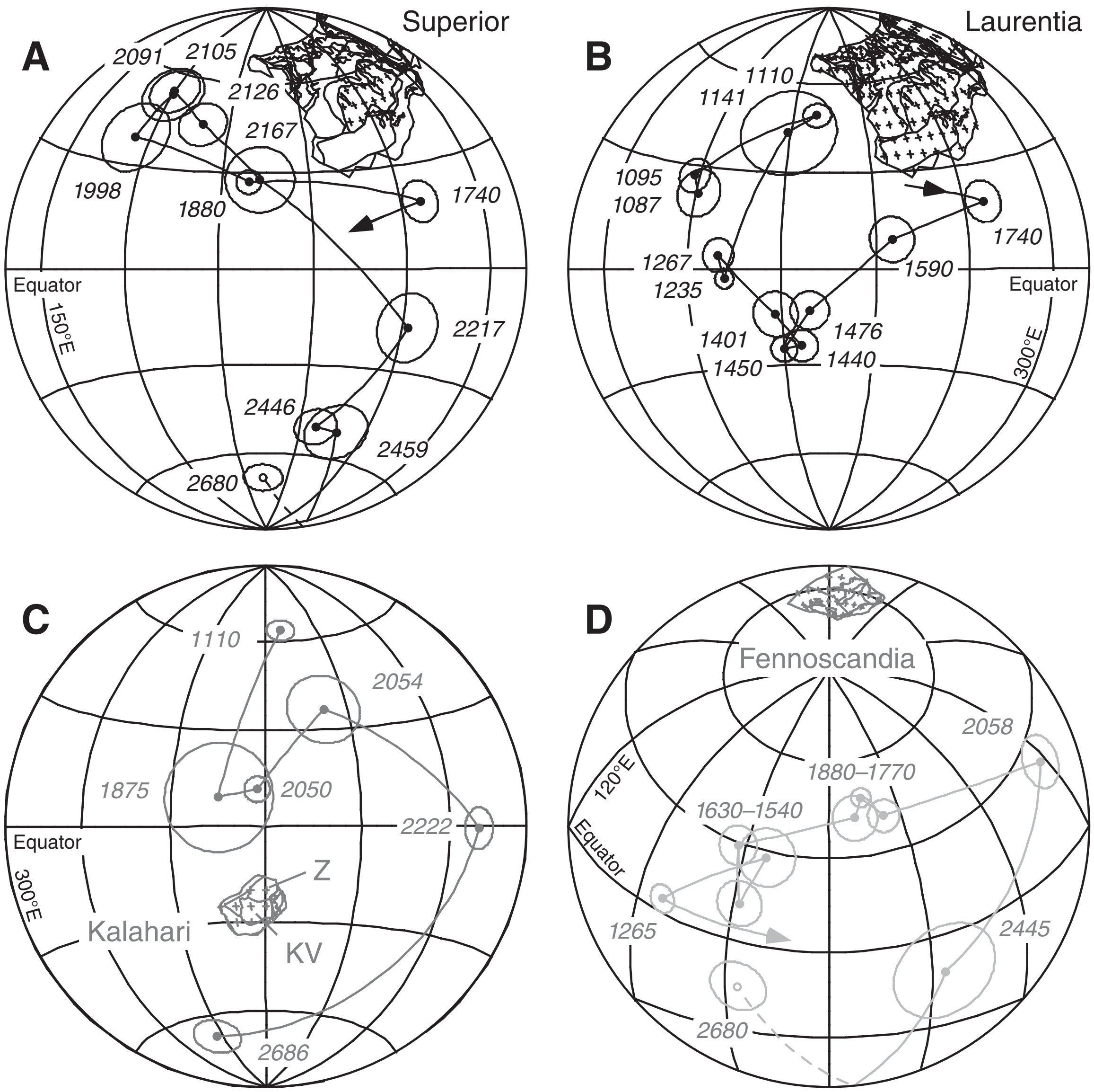

Figure 2. Paleomagnetic poles amenable to precise apparent polar wander (APW) length tests between pre–800 Ma cratons, in each craton's present coordinate reference frame. Greater complexities in the APW paths are evident from a broader consideration of non-key poles, but these are omitted here for clarity. As in subsequent global equal-area projection figures, black color represents Laurentia or the Superior craton, whereas gray color represents other cratons; all ages are in Ma. (A) Superior APW path. (B) Laurentia APW path, continuing in age progression from the Superior path. (C) Kaapvaal–Kalahari APW path (KV—Kaapvaal craton area; Z—Zimbabwe craton area). (D) Karelia–Fennoscandia APW path. All poles are listed in Tables 1 and 2, with the exception of the 1880–1540 Ma Fennoscandian poles (non-key; listed in Buchan et al., 2000; or Pesonen et al., 2003), which are included for comparison with Figure 5.

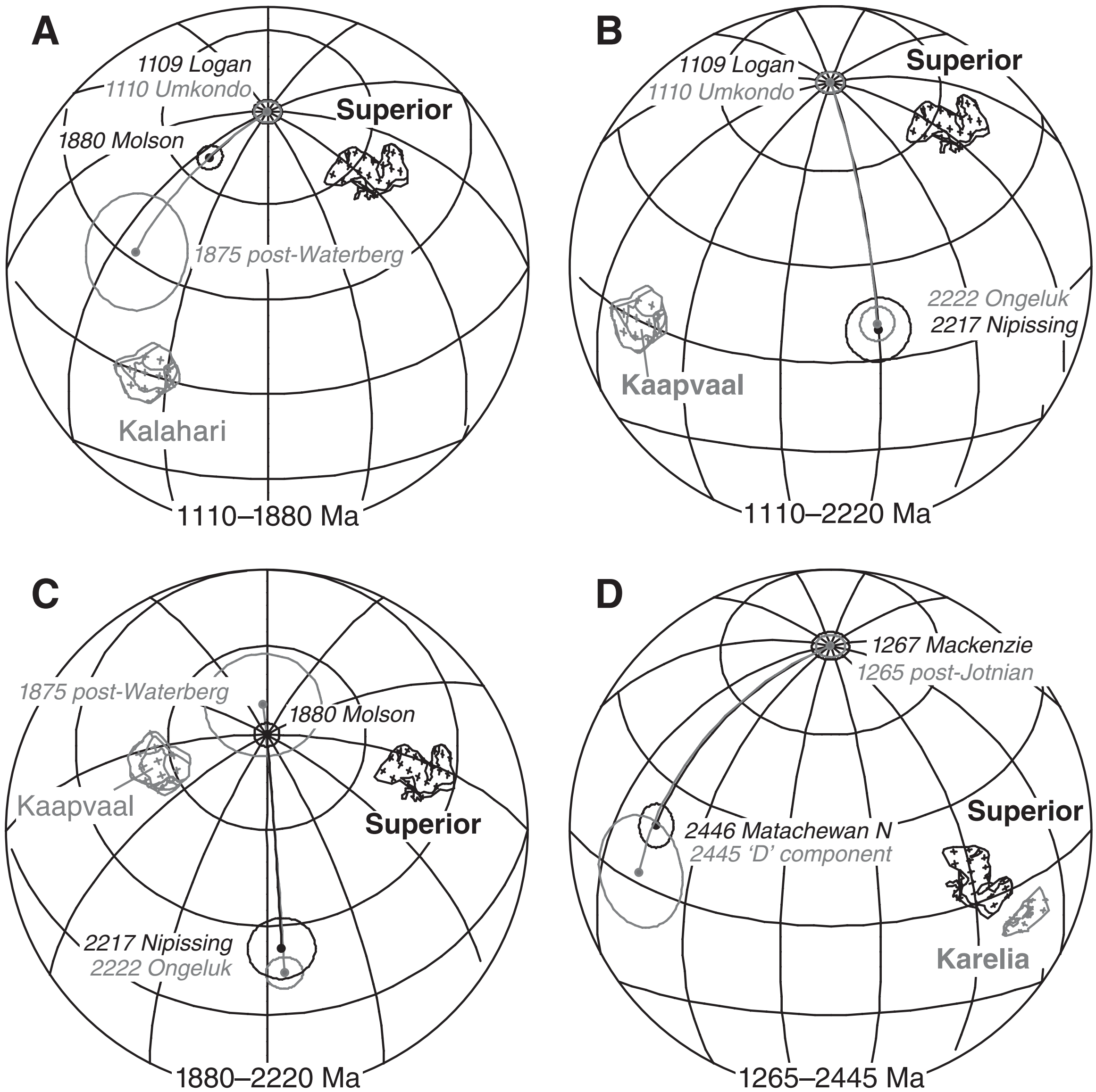

Figure 3. Reconstructions of cratons according to the pairs of key poles identified in Table 2. Due to polarity uncertainties, these reconstructions are nonunique and represent only half of the viable solutions for each pole pair. Coordinate system is the reconstructed paleolatitude grid of the younger pole in each pair. (A) Superior-Kalahari, 1110–1880 Ma; (B) Superior-Kaapvaal, 1110–2220 Ma; (C) Superior-Kaapvaal, 1880–2220 Ma; (D) Superior-Karelia, 1265–2445 Ma. Note that only the first panel (A) illustrates demonstrably independent motion between those two cratons (indicated by the different apparent polar wander [APW] lengths).

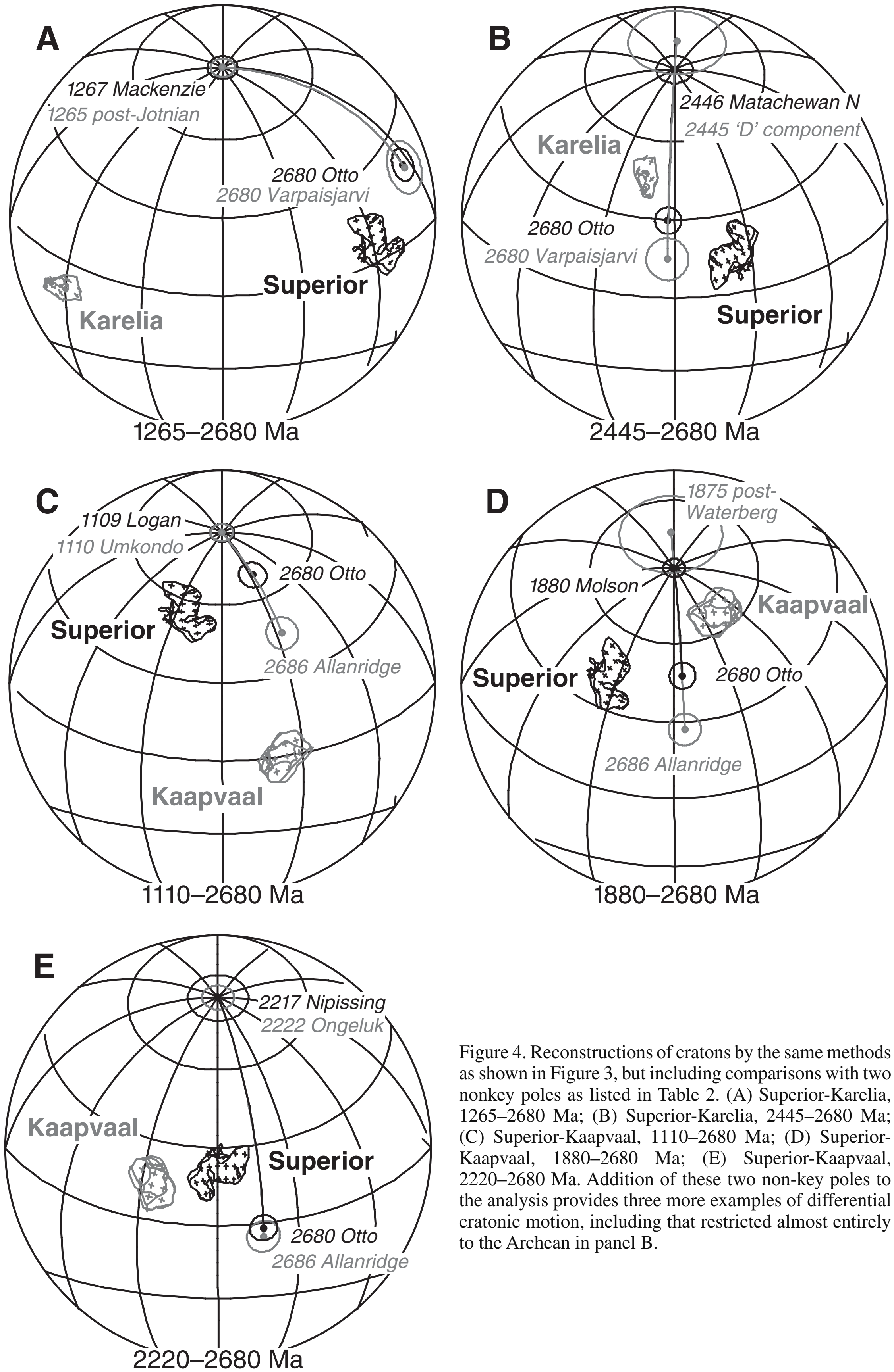

Figure 4. Reconstructions of cratons by the same methods as shown in Figure 3, but including comparisons with two nonkey poles as listed in Table 2. (A) Superior-Karelia, 1265–2680 Ma; (B) Superior-Karelia, 2445–2680 Ma; (C) Superior-Kaapvaal, 1110–2680 Ma; (D) Superior-Kaapvaal, 1880–2680 Ma; (E) Superior-Kaapvaal, 2220–2680 Ma. Addition of these two non-key poles to the analysis provides three more examples of differential cratonic motion, including that restricted almost entirely to the Archean in panel B.

similar to that of the Otto Stock as described already. Because of the three key pole comparisons between Superior/Laurentia and Kaapvaal/Kalahari already noted, addition of this single fourth pole from Africa, in comparison with the Otto Stock result from Superior, provides three new APW length tests. Shown in Figure 4 (panels C, D, and E), the younger two tests are positive, requiring differential motion between the cratons, whereas the oldest test is negative. In that last example (Fig. 4E), an intriguing juxtaposition is presented that could indicate a supercraton connection between the two blocks (sensu Bleeker, 2003), as first suggested by Evans and Powell (2000). Although the test is negative in terms of identifying differential cratonic motion and early plate tectonics on Earth, it could instead lead the way toward insights into Archean-Paleoproterozoic paleogeography.

Recognizing that these two examples are not yet of high reliability, we discuss them merely to demonstrate how tractable the APW length-comparison method can be: a few well-chosen subject rocks for paleomagnetic study in the near future may provide convincing evidence for Archean differential motions between cratons, or alternatively, produce potential supercraton juxtapositions of similar antiquity.

Our analysis using APW lengths has been highly conservative, in that we have selected only the most reliable paleomagnetic data. There are many published results that we suspect indicate primary and accurate magnetizations but that cannot be included in this study due to the strict quality criteria outlined already. If we were to include the entire database of moderately reliable results, we would certainly be reporting many additional examples of between-craton motions. Cawood et al. (2006) considered some of these and concluded that differential cratonic motion was demonstrable as old as 2070 and 2680 Ma. Those constituent data, however, suffer from uncertainties of paleohorizontal (e.g., the Kaapvaal ca. 2680 Ma pole from Mbabane pluton) or not precisely equal ages (e.g., ca. 2060–2050 Ma slowly cooling Bushveld complex versus the ca. 2075 Ma Fort Frances and slightly older dikes).

The second kind of paleomagnetic test we employed to determine possible differential cratonic motions utilizes known or hypothesized paleogeographic reconstructions and the logic developed in the discussion of Figure 1. If two cratons can be demonstrated to be joined in a supercontinent for a certain interval of time, then discordant older paleomagnetic poles in that reference frame will indicate differential cratonic motions in assemblage of the supercontinent. This type of test has been attempted in previous studies, although not using the precisely age-comparable and reliable subset of data selected herein. One example is the conclusion by Meert et al. (1995), based on discrepant ca. 800–600 Ma poles and well supported by tectonic syntheses (e.g., Collins and Pisarevsky, 2005), that Gondwanaland did not assemble until after that time. Another is the less confident hypothesis of Elming et al. (2001) that Baltica was not yet assembled at 1750 Ma, based on discrepant results from Sarmatia versus Fennoscandia. The tectonic assembly of Laurentia, better studied than any other Paleoproterozoic craton (Hoffman, 1988, and hundreds of subsequent publications), is a prime target for paleomagnetic analysis, and indeed, alternating fixist versus mobilist interpretations have been made through the past few decades (e.g., Christie et al., 1975; Cavanaugh and Nairn, 1980; Dunsmore and Symons, 1990; Irving et al., 2004; Symons and Harris, 2005). Recent work in western Canada, as yet published only in abstract (Buchan et al., 2007b; Evans and Raub, 2007) promises to clarify this debate with several new, key paleomagnetic poles. Initial results from those studies suggest that differential motion of Laurentian cratons can be documented between ca. 1880 and 1750 Ma. A final example, subtle and elegant, concerns a minor (10°–20°) vertical-axis rotation between the western and eastern halves of the Superior craton, as documented by dike trends and paleomagnetic data older than ca. 2000 Ma (Halls and Davis, 2004; Buchan et al., 2007a). The Kapuskasing zone serves as the locus of this deformation, which is constrained in age to ca. 1900–2000 Ma (Halls and Davis, 2004; Buchan et al., 2007a). Differential motion between (mostly) rigid blocks with narrow intervening boundary zones of deformation could well be considered as *bona fide* plate tectonics, but the continuation of Archean greenstone belts across the Kapuskasing zone implies that the total strain was minor (West and Ernst, 1991). The Superior example thus demonstrates a rheological form of plate-like tectonics, but not on a global scale.

An additional constraint may be provided by a particularly favorable cratonic reconstruction between Laurentia and Baltica for ca. 1265–1750 Ma. This reconstruction, named NENA by Gower et al. (1990), has been confirmed paleomagnetically by directional data from the Mackenzie and post-Jotnian mafic large igneous provinces (Buchan et al., 2000; Pesonen et al., 2003), as well as orientations of maximum magnetic susceptibility axes in rocks of those suites (Elming and Mattsson, 2001). Paleomagnetic results of lesser reliability, from rocks as old as ca. 1830 Ma, also support this connection (Buchan et al., 2000). The 1880 Ma poles from Superior and Fennoscandia, however, are discordant if rotated into the NENA reconstruction (Fig. 5). If NENA is valid to ages as old as ca. 1830, as supported by both paleomagnetic and tectonostratigraphic data (Gower et al., 1990; Buchan et al., 2000), then these data indicate relative motion between Superior and Karelia to achieve the assembly between 1880 and 1830 Ma. Unfortunately, the 1830 Ma poles are not reliable enough to support this conclusion definitively, but future work on rocks of this age has the potential to strengthen the argument considerably.

The demise of NENA is indicated by diverging APW paths of Laurentia and Baltica for ages younger than 1265 Ma. There is no precise estimate of the separation age; the oldest discrepant poles are ca. 1050 Ma, and those are not of “key” status (Fig. 5; poles are discussed in Buchan et al., 2000). Nonetheless, the overall APW pattern is intriguingly similar to that predicted by a plate-tectonic model of supercontinental assembly and dispersal, as illustrated in Figure 1. Concordance of the paleomagnetic reconstruction with a tectonically based juxtaposition (Gower et al., 1990) is the most compelling attribute of the aggregate

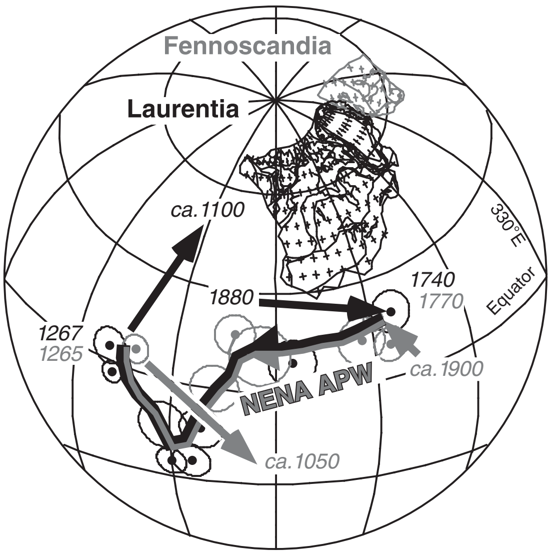

Figure 5. 1880–1100 Ma key paleomagnetic poles from Laurentia/Superior (black) in its present coordinate reference frame, compared with mostly non-key Baltica/Fennoscandia poles (gray) rotated to a similar configuration as "NENA" by Gower et al. (1990). Rotation parameters: (47.5°, 001.5°, +49.0°). References are in Table 1, or in Buchan et al. (2000), or Pesonen et al. (2003). Apparent polar wander (APW) paths converge between 1880 and ca. 1750 Ma, during the time of NENA assembly, and they diverge after 1265 Ma, coincident with NENA breakup.

data set. According to this evidence, modern-style plate tectonics appears to have been in operation throughout the most of the Proterozoic, negating the Mesoproterozoic stagnant-lid model of Stern (this volume).

CONCLUSIONS AND FUTURE PROSPECTS

Limitations of the Archean geological record, particularly the dearth of subgreenschist stratified volcano-sedimentary successions older than 3000 Ma—as well as our currently poor knowledge of the Archean geomagnetic field—substantially hinder paleomagnetic methods to test whether plate tectonics existed at those times. The available highest-quality paleomagnetic data provide only a single example of differential cratonic motions within the Paleo- and Mesoproterozoic Eras, when at least some portions of Earth's surface were rigid blocks moving laterally relative to each other. Lithospheric rigidity and lateral motion are two important aspects of a tectonic regime that is "plate-like," and, therefore, the most robust paleomagnetic record argues for at least one instance of plate tectonics prior to 1110 Ma (Table 2).

We noted herein that if certain "non-key" poles were used in the analysis, then the differential motions could be better constrained, by various paleomagnetic methods, to as old as the earliest Proterozoic Eon. In the longer term, the present analysis might be able to extend Earth's earliest record of greenschist- or lower-grade stratified rocks, primarily known from the Barberton area of Kaapvaal in southern Africa, and the Pilbara region of Western Australia. These two terranes have been hypothesized to constitute our planet's oldest "supercontinent" (Zegers et al., 1998; extending the work of Cheney, 1996), but paleomagnetic studies of stratified rocks from the 3500–2800 Ma interval are few in number (e.g., McElhinny and Senanayake, 1980; Yoshihara and Hamano, 2004), and only one result satisfied our quality criteria. Even that lone result, the magnetite-borne remanence from the Duffer Formation in the Pilbara craton (McElhinny and Senanayake, 1980), is only constrained to be prefolding in age, and the folding could have occurred more than 100 m.y. following deposition (Van Kranendonk et al., 2002). Establishing the 700-m.y.-long APW paths for the Kaapvaal and Pilbara cratons will be challenging at the least, and might even be impossible given the moderate metamorphic grade and difficulties in testing whether Earth's primitive geodynamo can serve as a stable kinematic reference frame. Nevertheless, these two successions hold the greatest promise for paleomagnetism to address whether plate tectonics existed deep within Archean time.

We showed here that incorporation of just two merely moderately reliable sets of paleomagnetic data allowed us to identify differential motion of cratons as old as 2445–2680 Ma. If there is interest within the paleomagnetic community, those and similarly ancient data can be "upgraded" to the highest level of quality within a matter of years. A more extensive research program on the 2800–3500 Ma volcano-sedimentary successions of Kaapvaal and Pilbara holds the best promise for detecting any relative motions between cratons in the older part of the Archean Eon.

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