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Chapter 2

Origin and Timing of Banded Iron Formation-Hosted High-Grade Hard Hematite Deposits—A Paleomagnetic Approach

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Abstract

The processes responsible for the transformation of banded iron formations to hard high-grade hematite ore, and their timing, remain poorly understood despite many recent advances. The paleomagnetic method allows for the estimation of ore genesis timing as a complement to other techniques. The effectiveness of the paleomagnetic method at dating, and testing proposed models for, the genesis of hard high-grade hematite ore deposits is illustrated by two South African examples. A new dataset is reported for the Thabazimbi deposit that independently constrains the age of ore formation between 2054 and 1930 Ma, while previously published data from the Sishen-Beeshoek deposits highlight the association of those deposits with weathering preceding the development of a marked Paleoproterozoic-aged unconformity (older than 2060 Ma). Paleomagnetic results are in both cases consistent with proposed models of ore genesis (i.e., extensive carbonate metasomatism and meteoric fluid interaction at Thabazimbi and ancient supergene processes at Sishen-Beeshoek). The antiquity of these South African examples appears to reflect a common theme among other hard high-grade hematite deposits from around the world, as revealed by a review and re-evaluation of existing paleomagnetic literature. This review represents a first attempt at providing a synopsis of hard high-grade hematite deposits within a temporal framework. The apparent Paleoproterozoic to Mesozoic age distribution of deposits as discussed in this review, which must be tested and verified by both the expansion of the database and improvement of current available data, has important implications for proposed models of ore genesis, as well as for exploration.

Introduction

HIGH-GRADE (60 wt % Fe) hematite iron ore deposits that are hosted by banded iron formation (BIF) are known for all the continents, excluding Antarctica. These deposits are the products of the enrichment of Precambrian BIFs, and they are one of the most important sources of iron ore today (U.S. Geological Survey, 2007). Many high-grade hematite orebodies consist of "soft" martite-hematite ore, which is generally accepted as having formed through recent supergene processes (Harmsworth et al., 1990). It is, however, the so-called "hard" high-grade hematite ores, which consist of martite and microplaty hematite, that are the focus of this study. The processes by which BIF is upgraded to hard high-grade hematite ore are still a matter of debate, and a variety of models have been suggested. Models range from syndepositional (King, 1989) or diagenetic (Findlay, 1994); synorogenic, where upgrading takes place via heated orogenically driven fluids (Li et al., 1993; Powell et al., 1999); modern supergene (Macleod, 1966) or ancient supergene (Morris, 1980, 1985; Beukes et al., 2002, 2003); to composites that invoke fluid mixing, for example, the models of Barley et al. (1999) and Taylor et al. (2001) At Carajas, Lobato et al. (2005) have implicated magmatic fluids related to deep fault-magmatic systems as being responsible for the upgrading of BIF to hard high-grade hematite ore (see also Lobato et al., 2008).

Despite recent progress made in elucidating the processes involved in the upgrading of BIF to ore (e.g., Webb et al., 2003; Thorne et al., 2004; Webb et al., 2004; Figueiredo e Silva et al., 2007; Oliver et al., 2007, among others), our knowledge of fluid sources and the timing of ore formation is scant. This is true for most of the hydrothermal deposit types (be they related to basinal brines, meteoric fluids, or magmatic fluids). There are, however, significant differences in what is known from different deposit types. In order to develop new exploration targets, it is of the utmost importance to understand the mechanisms of ore genesis, in particular the age of controlling structures and of the sources and timing of fluids that govern ore formation.

Fine-grained hematite can be a carrier of very stable remanent magnetization (Dunlop and Özdemir, 1997), and different generations or grain-size fractions of hematite may record different magnetizations that reflect the ambient magnetic field at the times of crystallization (Stokking and Tauxe, 1990). It follows that high-grade hematite deposits, and perhaps

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even distinct phases of multistage ore formation, should be datable by the paleomagnetic method (i.e., by comparing the paleomagnetic pole obtained from the ore with the apparent polar wander path (APWP) of the relevant continent or crustal block). Such relative dating permits crude estimation of age to complement other methods. It is particularly suited to constrain the timing of mineralization and of tectonic events. The use of APWP to date ore deposits has been applied successfully to Mississippi Valley-type (MVT) lead-zinc deposits (e.g., Leach et al., 2001) and some of the world's largest manganese deposits (Evans et al., 2001), but the method has had only limited success at dating high-grade hematite deposits. The reasons for this are set out in the review and conclusion part of this paper.

In this contribution we attempt to test the models proposed for, and timing of, two South African high-grade hematite deposits (summarized by Beukes et al., 2003): the Thabazimbi deposit, by means of a new paleomagnetic dataset; and the Sishen-Beeshoek deposits, from previously published data (Evans et al., 2002). In addition, a review of paleomagnetic studies conducted on hard high-grade hematite deposits is presented. To identify potential fluid sources, the paleomagnetic age constraints are compared with the geologic record in order to identify any correspondence with marked tectonic, erosional, or magmatic events. This paper presents the first such attempt at providing a synopsis of hard high-grade hematite deposits within a temporal framework.

Origin of the Thabazimbi Deposit, South Africa

The Thabazimbi deposit (Fig. 1) is a typical example of an economically important BIF-hosted high-grade hematite deposit for which the origin has remained unresolved. Ore-forming processes invoked by previous authors range from supergene (Wagener, 1921; Boardman, 1948) and magmatic (De Villiers, 1944) to metasomatic-hydrothermal with a later supergene overprint (Du Preez, 1944; Strauss, 1964; Van Deventer et al., 1986). Advances in the understanding of the origin of the Thabazimbi deposit were recently reported by Gutzmer et al. (2005). Their findings and the general geologic setting of the Thabazimbi deposit are summarized here.

The Thabazimbi deposit is composed of a series of strata-bound orebodies that are aligned along a faulted basal contact between the 2.48 to 2.43 Ga Penge Iron Formation (Trendall et al., 1990; Walraven and Martini, 1995) and dolostone of the 2.58 to 2.52 Ga Malmani Subgroup (Walraven and Martini, 1995), Transvaal Supergroup (Fig. 1). Orebodies further appear to be duplicated by a series of east-west–striking thrusts.

In the area surrounding the Thabazimbi deposit the Transvaal Supergroup strata dip steeply below the Bushveld Complex, which intruded the Transvaal Supergroup at ca. 2060 Ma (e.g., Buick et al., 2001). The presence of pseudomorphs of microplaty hematite after grunerite and ankerite suggests that ore formation postdated peak contact metamorphism associated with the intrusion of the Bushveld Complex (Gutzmer et al., 2005). The tilting of the Transvaal Supergroup is the result of the thermal adjustment of the crust following the intrusion of the Bushveld Complex (Eriksson et al., 1995). Thermal adjustment of the crust was also responsible for warping and faulting of the 2054 ± 4 Ma lowest unconformity-bounded sequence of the Waterberg Group (Dorland et al., 2006). The 2054 ± 4 Ma age (Dorland et al., 2006) from quartz porphyry lava at the base of the Waterberg Group thus provides a maximum age for tilting of strata following the intrusion of the Bushveld. There is no relative age constraint between high-grade hematite ore genesis and tilting.

Strata of the Transvaal Supergroup, high-grade hematite ores and granite of the Bushveld Complex are duplicated by thrust faults in the Thabazimbi area (Fig. 1), defining mountain ranges that strike approximately east-west. These prominent thrust faults postdate an event of tilting of the strata as well as an event of normal faulting, which is expressed by loss of stratigraphy along the contact between the Penge Iron Formation and the underlying dolostone of the Malmani

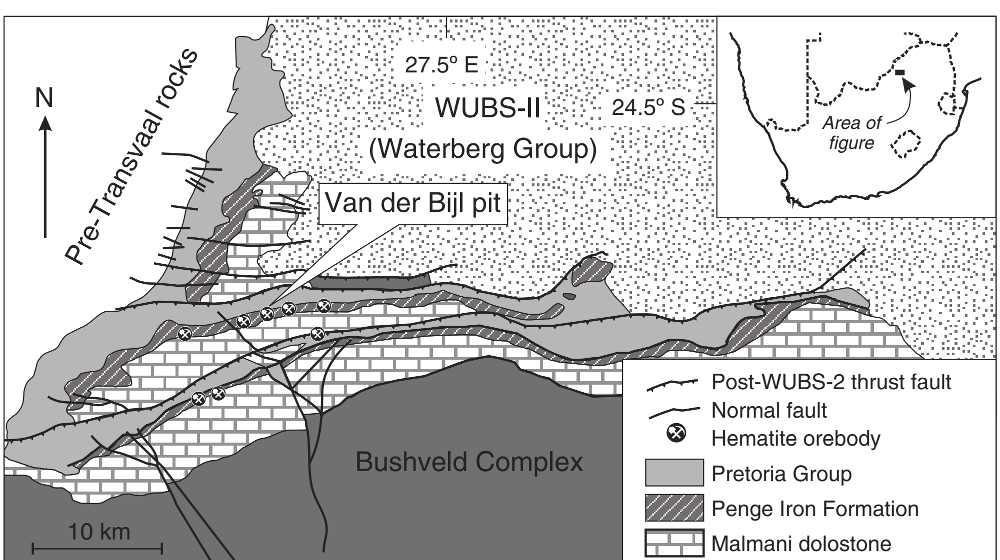

FIG. 1. Simplified geology of the Thabazimbi area. The location of hematite orebodies and the Van der Bijl deposit (sampled for paleomagnetic study) are indicated.

Subgroup (Gutzmer et al., 2005). The thrust fault system, and folding associated with it, is regarded to be post-Bushveld in age (Du Preez, 1944; Du Plessis, 1987; Gutzmer et al., 2005) as thrusts caused strata of the Transvaal Supergroup to override Bushveld-aged granite as well as Paleoproterozoic Waterberg red beds to the north of Thabazimbi (Fig. 1).

The red beds of the Waterberg Group have been divided into several unconformity-bounded sequences by Cheney and Twist (1986). The thrust faults affect rocks as young as the second unconformity-bounded sequence of the Waterberg Group, or WUBS-2 (e.g., Du Preez, 1944). Post-WUBS-2 thrusts, and related folds, appear to be truncated by the Sandriviersberg Formation of the sequence WUBS-3 of the Waterberg Group. Dorland et al. (2006) proposed that the WUBS-3 was deposited before 1930 Ma, based upon regional correlations with formations dated by zircon evaporation U-Pb techniques (Cornell et al., 1998). This could provide a minimum age for the Thabazimbi ore formation, but it is somewhat open to interpretation because the thrust faults actually die out within the WUBS-2 (Fig.1), and their truncation by the Sandriviersberg Formation is not directly observed (Du Plessis, 1987).

The Thabazimbi deposit is marked by the predominance of hard high-grade hematite ores. At depth the high-grade hematite ores interfinger with low-grade dolomite-hematite and calcite-hematite ores. Iron oxides (martite and microplaty hematite) present in these low-grade ores are texturally indistinguishable from those in the associated high-grade ores (i.e., ore formation appears to be genetically linked to hydrothermal carbonate metasomatism; Gutzmer et al., 2005). Microthermometric study of fluid inclusions in sparry carbonates in the low-grade carbonate-rich ores and rare megaquartz within the high-grade ores reveals their shared hydrothermal origin (Gutzmer et al., 2005). This suggests that the hard high-grade hematite orebodies at the Thabazimbi deposit owe their origin to an event of extensive oxidative carbonate metasomatism, perhaps similar to that recognized in the Hamersley district in Western Australia (Webb et al., 2004).

Hydrothermal ore formation at Thabazimbi is both structurally and lithologically controlled (Fig. 1). The size and quality of the orebodies were improved by late Mesozoic-Cenozoic deep lateritic weathering, as carbonate-hematite ores were enriched to form friable or soft high-grade hematite ores. This supergene modification is, according to Gutzmer et al. (2005), secondary in importance compared to the primary event of hypogene metasomatic ore formation. While it is clear that ore formation followed the intrusion of the Bushveld Complex, the relationship between ore formation and the tilting of strata is more ambigious.

The Sishen-Beeshoek Deposits, South Africa

The Sishen and Beeshoek deposits are large, hard high-grade hematite deposits in the northern Cape Province of South Africa and considered type examples of ancient supergene deposits (Beukes et al., 2003). The deposits are situated near the western edge of the so-called Maremane dome (Fig. 2), an anticlinal structure with a core of dolostone overlain by BIF of the Asbesheuwels Subgroup. On the western edge of the Maremane dome (along the Gamagara ridge) and along the so-called Klipfontein hills to the east, scattered outcrops of chert-breccia and BIF are present. The hematite ore deposits of the Sishen and Beeshoek deposits are hosted within this BIF, named the Manganore Iron Formation, which is regarded as an altered or oxidized equivalent of the Kuruman Iron Formation and Griquatown Iron Formation of the Asbesheuwels Subgroup (Beukes and Smit, 1987). The Manganore Iron Formation, and succeeding diamictite, andesitic lava, and chemical sedimentary rocks belonging to the Postmasburg Group, are unconformably overlain by a Paleoproterozoic red-bed succession containing the Mapedi and Gamagara Formations (Beukes and Smit, 1987; Beukes et al., 2002). The hard high-grade hematite ore at Sishen and Beeshoek is developed directly below the unconformity at the base of the Mapedi-Gamagara unit (Fig. 2). Large deposits of high-grade hematite deposits are developed exclusively where the iron formation has slumped via paleokarst into dolostones of the underlying Cambellrand Subgroup along the Maramane dome (Beukes et al., 2003). Weathering along the base of the Mapedi-Gamagara unit and the associated upgrading of BIF are estimated as having occurred around 2.18 to 2.22 Ga (Beukes et al., 2002).

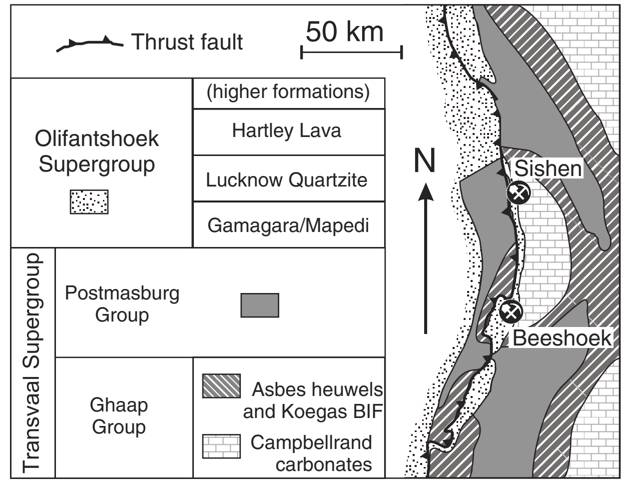

FIG. 2. General geology and stratigraphic setting of the Griqualand West basin (after Evans et al., 2002).

The proposed models of origin and timing of mineralization at the Thabazimbi and Sishen-Beeshoek deposits differ significantly from each other, and since the Paleoproterozoic APWP of the Kaapvaal craton is fairly well constrained (de Kock et al., 2006), these deposits provide an excellent opportunity for illustrating the power of the paleomagnetic method for indirectly dating the mineralizing events.

Paleomagnetic Method

Fourteen individually oriented drill core samples of hard high-grade hematite ore were collected from the Van der Bijl pit (Fig. 1) of the Thabazimbi deposit for paleomagnetic study and comparison with the published Paleoproterozoic APWP of the Kaapvaal craton (de Kock et al., 2006). Trimmed core samples from Thabazimbi were treated to sequential alternating-field (AF) and thermal demagnetization and measured with a 2G-Enterprises™ SQuID magnetometer that is housed at the California Institute of Technology in a magnetically shielded chamber. A typical demagnetization sequence started with measurement of natural remanent magnetization (NRM), followed by low field strength AF pretreatment in five successive steps to 10 mT. Samples were then thermally demagnetized in a shielded furnace at decreasing intervals until specimen intensity dropped below noise level, or more typically in the present study, magnetizations became erratic due to acquisition of spurious directions at the highest levels of thermal demagnetization, above 680°C (e.g., Evans et al., 2002). Magnetic components were identified and quantified via least squares principal component analysis (Kirschvink, 1980). All calculations of virtual geomagnetic poles and paleomagnetic poles assume an axial-geocentric dipolar magnetic field and a paleoradius for the Earth equal to the present Earth radius.

Poles obtained from high-grade hematite ores in South Africa and around the world were compared to various relevant APWP, for example, the suggested Paleoproterozoic APWP of the Kaapvaal craton in the case of the Thabazimbi and Sishen-Beeshoek deposits. When a continuous APWP has not been suggested for a specific time interval, representative and reliable poles were selected from the global paleomagnetic database (GPMDB v 4.6) for comparison. These poles were selected by using the reliability criteria of Van der Voo (1990). During the selection process poles were included only if they are well dated (with errors of ± 4%) and if a presumption of primary magnetization is allowable. In addition, poles must be derived from studies that employed adequate demagnetization and that satisfy basic statistical criteria (k >10 and α_{95} <16°). In extreme cases where paleomagnetic data are scarce, some well-dated poles that are interpreted as regional overprints were also included. They are clearly identified when used.

The poles from high-grade hematite ores are in most cases poorly constrained (as described in the review section below) and commonly do not meet the same reliability criteria as that of the APWP with which they are being compared. The criteria involved in pole selection, and the vigorous process usually associated with the definition of APWP, are thus seen as prerequisites for ensuring the best possible age estimates of ore genesis. In addition, for the purposes of this paper, the poles from high-grade hematite ores are assumed to be representative of some phase of ore genesis or hematite formation. It is acknowledged that ores may have been recrystallized or altered at some point in time, and it is possible that no paleomagnetic record remains of the initial phases of ore genesis. The ages obtained from the paleomagnetic method are thus considered minimum estimates for the onset of ore-forming events.

Paleomagnetic Results from Thabazimbi

Ten of the 14 collected samples contained two magnetic components, whereas four recorded only single components. The reproducibility of sample behavior could be tested in cases where two "sister specimens" were trimmed from the same core. Sister specimens displayed nearly identical behavior.

At lower demagnetization levels (<550°C) samples displayed linear demagnetization trajectories (Fig. 3), but the overall grouping of these directions is scattered (Fig. 4) and no mean was calculated. With continued demagnetization between about 600° and 680°C, trajectories remained fixed at stable end points (Fig. 3). At higher temperatures sample behavior generally became erratic. Linear connections between stable end points and the origin define southerly and moderate-steep upward magnetic components (Fig. 3). No high-temperature components could be calculated for samples TV 1 and TV 11 due to erratic behavior of these samples at temperatures in excess of ~550°C. The behavior of sample TV 8 differs from the rest of the samples in that its high-temperature component is northwesterly directed (Fig. 4) but with a similar moderate-steep inclination as the rest of the samples. It is a clear outlier, and overprinting by the present-day geomagnetic field (i.e., weathering) is suspected. Sample TV 14 is inclined much shallower than the rest of the samples and is directed southeasterly rather than southerly (Fig. 4). Exclusion of this sample significantly improves the high-temperature component mean (n = 10, declination = 176.1°, inclination = –58.9°, k = 16.0, α_{95} = 11.8°). A structural correction for the tilt of the orebody (parallel to strata exposed within the mine; strike/dip 102°/79° S) shifts the mean to a north-northeasterly position (declination = 23.1°, inclination = –40.6°). Note that orientation of strata within the Van der Bijl pit differs slightly from the regional strike of the Transvaal Supergroup in the area, which is east-northeast. Paleomagnetic poles were calculated for both geographic and structurally corrected means. These are situated at 25.6° N, 024.1° E (dp = 13.1°, dm = 17.5°) and 68.9°S, 298.4° E (dp = 8.4°, dm = 14.2°), respectively, after inverting the polarities of both directions for comparison with the APWP segments illustrated in Figure 5.

Compared to Paleoproterozoic paleomagnetic poles of the Kaapvaal craton (de Kock et al., 2006), it is the structurally uncorrected pole from the Thabazimbi orebody that stands out as resembling several of the previously reported poles (Fig. 5A; Table 1). The structurally corrected Thabazimbi hard ore pole shares some similarity with the Mesoproterozoic APWP of the Kalahari craton (Fig. 5B), but this coincidental similarity is refuted by the stratigraphic-structural evidence for the timing of ore formation (i.e., the ore probably formed prior to thrusting of the Transvaal Supergroup and lower Waterberg Group, as described above). The uncorrected Thabazimbi orebody pole plots slightly southwest of the WUBS-1 pole from the Lower Waterberg Group and the 2060 Ma Phalaborwa Complex (PB1) pole. It furthermore displays good overlap with the Vredefort impact structure virtual geomagnetic pole, which is dated at ca. 2020 Ma (Kamo et al., 1996; Gibson et al., 1997), as well as with the less well-dated Bushveld Main and Upper zones mean cooling pole compiled by Evans et al. (2002). Due to a loop in the Paleoproterozoic APWP of the Kaapvaal craton, a fair degree of overlap is also shared with the ca. 1875 Ma pole for post-Waterberg sills (Hanson et al., 2004; pole PWD in Fig. 5), which has been recalculated to exclude the demonstrably younger sills and lavas of the Soutpansberg Group (de Kock et al., 2006; Dorland et al., 2006).

Petrographic constraints from the Thabazimbi region suggest that high-grade hematite ore formed after Bushveld magmatism at 2060 Ma. Paleomagnetic data are in good agreement with this, as the Thabazimbi orebody pole plots on a segment of the craton's APWP that postdates the 2054 Ma WUBS-1 pole (Fig. 5A). Our paleomagnetic data also makes it now clear that ore formation took place after tilting of the strata. The stratigraphic relationships provided by the apparent duplication of the orebody by pre-WUBS-3 thrust faults further constrain the Thabazimbi orebody pole on the APWP

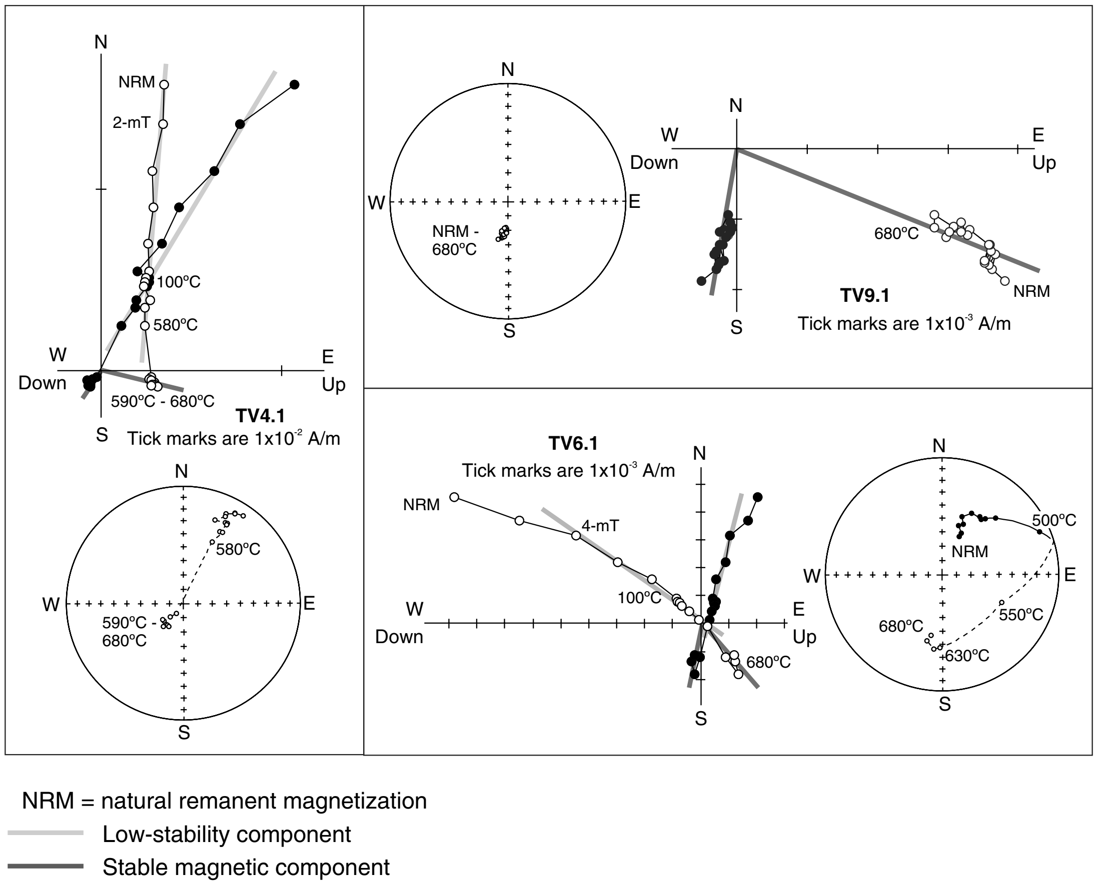

FIG. 3. Demagnetization behavior of representative samples of hard hematite ore from the Van der Bijl deposit in in situ coordinates. Orthogonal plots: solid symbols = horizontal plane, open symbols = north-south or vertical plane. Equal area plots: solid symbols = upper hemisphere, open symbols = lower hemisphere.

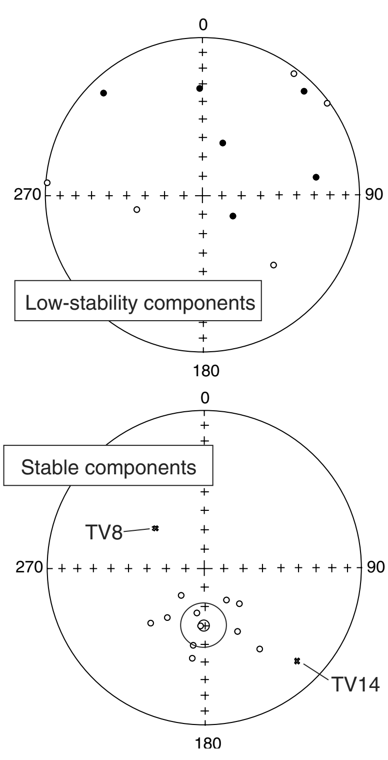

FIG. 4. Equal area plots of least squares analysis results for low stability and stable components. Solid symbols = upper hemisphere, open symbols = lower hemisphere. Oval is cone of 95 percent confidence about the mean calculated from stable components. The mean is represented by the larger open symbol. Crosses represent outliers and possible orientation errors.

segment that represents the timing of deposition of the lower two unconformity-bounded sequences of the Waterberg Group (2.05–1.9 Ga). On this segment of the APWP the Thabazimbi orebody pole plots closer to the WUBS-1 pole (Fig. 5A), thus suggesting that ore formation occurred closer in time to the deposition of the lowermost Waterberg Group. A recent post-Bushveld 2042 ± 3 Ma $^{40}Ar/^{39}Ar$ age for low-grade metamorphism and deformation in the Transvaal Supergroup (Alexandre et al., 2006) may date the deformation associated with the thermal relaxation of the central Kaapvaal craton following the intrusion of the Bushveld Complex. This age is assigned with caution as an approximate age for the structurally uncorrected Thabazimbi orebody pole.

We note the possibility that thrust faults deforming both the orebody and sequence WUBS-2 could in fact be younger than sequence WUBS-3. In that case the ore could have formed at ca. 1875 Ma, as allowed by similarity between the Thabazimbi orebody and PWD poles (Fig. 5A).

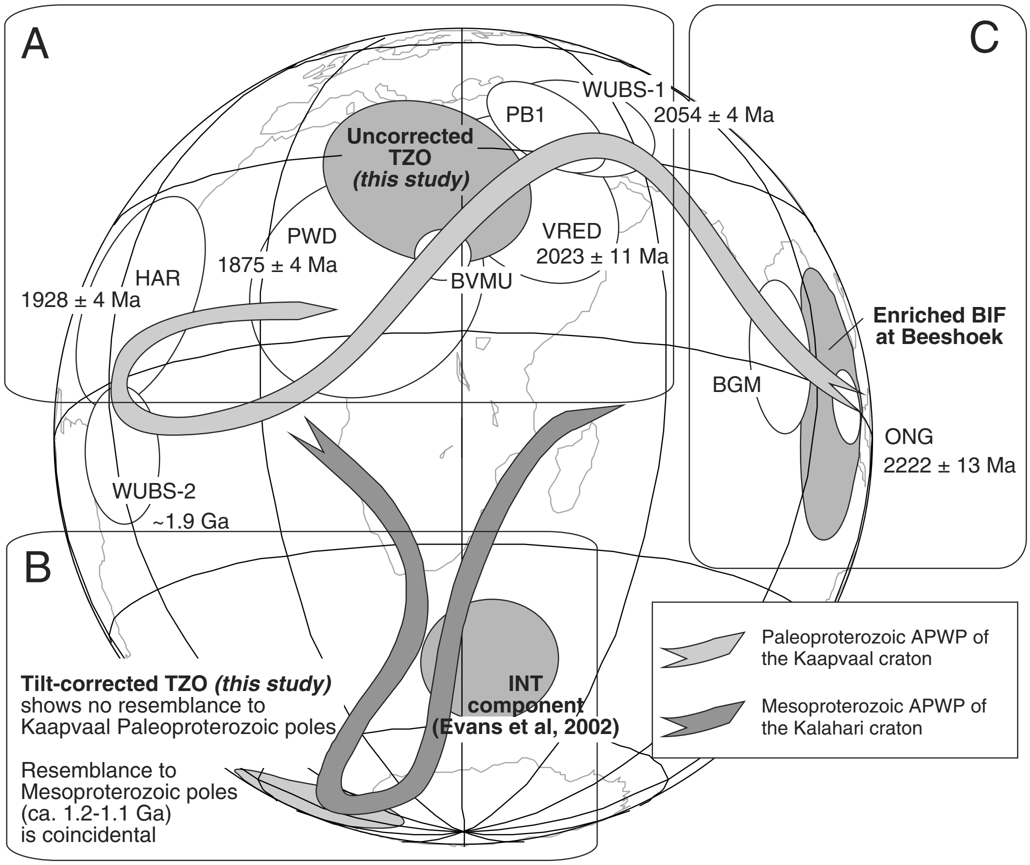

FIG. 5. A. View from space of the virtual geomagnetic pole obtained for the magnetically stable component (in situ) recorded by hard hematite ore at Thabazimbi. Also shown are some of the known Paleoproterozoic poles of the Kaapvaal craton (listed in Table 1) as well as the Paleoproterozoic APWP of the Kaapvaal craton (stylized after De Kock et al., 2006). B. The structurally corrected pole for the hard ore at Thabazimbi together with the Mesoproterozoic APWP of the Kaapvaal craton (stylized after Evans et al., 2002). This view also shows the position of the INT pole recorded at Beeshoek, overlapping with that of the Namaqua metamorphic province. C. Pole for enriched BIF from Beeshoek, showing proximity to the basal Gamagara and/or Mapedi Formation poles.

The ore formation at Thabazimbi is both structurally and lithologically controlled (Gutzmer et al., 2005). A Paleoproterozoic age for the Thabazimbi deposit, as confirmed by our paleomagnetic results, suggests that thrust faults may have brought up the Penge Iron Formation at Thabazimbi to a paleoweathering horizon just below WUBS-III. Isotopic and fluid inclusion studies (Gutzmer et al., 2005), however, suggest a nonsupergene origin for the ores. Paleomagnetic study of the carbonate-hematite protore at Thabazimbi is needed to confirm the coeval nature of the protore and hard high-grade hematite as suggested by Gutzmer et al (2005).

Discussion of Paleomagnetic Results from Sishen-Beeshoek

Evans et al. (2002) conducted a detailed paleomagnetic study of the basal part of the Gamagara Formation and its lateritized substrate with the goal of constraining the age of lateritic weathering. As part of their study they sampled hard high-grade hematite ore from the north Sishen mine, enriched BIF from the Manganore iron ore deposit at Beeshoek, and hematitic conglomerates with enriched ore pebbles from both localities. In this paper we summarize their results and the implications thereof for the proposed timing of mineralization.

A conglomerate test is a field stability test for the age of a paleomagnetic remanence direction and is positive when individual clasts within a conglomerate display uniformly distributed or "random" magnetic directions as opposed to well-grouped magnetic components for the parent lithology from which the clasts were derived. A positive conglomerate test constrains the timing of magnetization to before the formation of the conglomerate.

Hard high-grade hematite ore samples from the north Sishen mine displayed stable end points of magnetization at 680ºC, but they were decidedly scattered, thus denying Evans et al. (2002) from any conclusive interpretation on an apparently positive conglomerate test from the same locality. Demagnetization of enriched BIF samples from Beeshoek, however, was extremely stable against thermal demagnetization and displayed single relatively well grouped magnetic components (named HIG- and referring to negative inclination components revealed at high-temperature demagnetization steps). Clasts from a hematite ore-pebble conglomerate from Beeshoek displayed randomly oriented components, in

TABLE 1. Paleopoles Used to Evaluate Age Assignments of High-Grade BIF-Hosted Hard Hematite Deposits

Rock unit	Abbreviation	Age and uncertainty	Age reference	PLAT	PLONG	DP	DM	Pmag reference	Q
West African craton (post-1.7 with the exclusion of L-MET, IC1, and IC2; see text)									
Edjeleh fold	EF	309.5 ± 2.5 Ma	1	28.3	238.9	4.2	4.2	2	6
Oubarakat and El-Adeb Larache Formations	OUB	315 ± 3 Ma	1	28.2	235.5	3.4	3.4	3	5
Char Group Unit I2	CHR	998 ± 32 Ma or ~1.2 Ga ?	4, 5	48.7	206.6	1.4	2.7	6	4
Metamorphic rocks (Liberia)	L-MET	2050 ± 6 Ma or 1930 ± 38 Ma ?	7, 8	17.5	269.1	13.0	13.0	7	3
TTG granite (Ivory Coast)	IC2	2085 ± 15 Ma	8	–82.0	292.0	13.0	13.0	9	3
Ferke granite (Ivory Coast)	IC1	~ 2000 Ma	9	–25.0	83.0	16.0	16.0	9	4
Gawler craton (all ages)									
Group GB dikes (overprint)	GB	1.6–1.5 Ga	10	22.8	266.4	11.3	11.3	11	2
Group GA dikes (overprint)	GA	1.6–1.5 Ga	10	61.4	230.8	8.6	8.6	11	2
Gawler Range Volcanics	GR	1592 ± 2 Ma or ovp.	10, 12	60.4	230.0	6.2	6.2	13	4 or 5
Pilbara craton (2.935 Ga to 542 Ma) including the post-1.8 Ga poles from the Yilgarn craton									
Millindinna Complex	MC	2860 ± 20 Ma	14	–11.9	161.3	6.8	8.4	14	5
Black Range dikes	BR	2772 ± 2 Ma	15	32.0	334.0	9.0	9.0	16	6
Cajuput dike	CP	2772 ± 2 Ma	15	46.0	326.0	22.0	22.0	16	5
Mount Roe Basalt	MR	2772 ± 2 Ma	16	–52.4	178.0	6.4	9.1	14	7
Pilbara flood basalts package 1	P1	2772 ± 2 Ma	15	40.8	339.8	3.7	3.7	18	6
Pilbara flood basalts package 2	P2	2768 ± 16 Ma	19	46.5	332.7	15.2	15.2	18	6
Pilbara flood basalts package 4-7	P4-7	2752–2715 Ma	17, 20	50.4	318.2	12.5	12.5	18	6
Mount Jope Volcanics	MJ	2721–2713 Ma	20	–40.5	128.7	19.9	20.8	14	4
Pilbara flood basalts package 8-10	P8-10	2721–2713 Ma	20	59.1	6.3	6.1	6.1	18	6
Brockman BIF at Paraburdoo	PBB	1750 ± 50 Ma or ~2.22–2.03 Ma	21, 22	36.4	29.6	4.7	8.8	23, 24	2
Brockman BIF at Witenoom	WM	1750 ± 50 Ma or ~2.22–2.03 Ma	21, 22	41.1	21.7	6	10.4	23, 24	2
Jeerinah Formation overprint	JRO	1750 ± 50 Ma or ~2.22–2.03 Ma	21, 22	41.6	32.2	4.5	8.5	23, 24	3
Wittenoom Dolomite overprint	WDO	1750 ± 50 Ma or ~2.22–2.03 Ma	21, 22	38.4	20.8	1.6	2.8	23, 24	3
Mt. Jope Volcanics overprint	JO	1750 ± 50 Ma or ~2.22–2.03 Ma	21, 22	35.0	31.5	2.9	5.5	14, 24	2
Pilbara flood basalts MT overprint	MT	1950 ± 150 Ma	18	53.2	23.9	5.3	5.3	18	3
Frere Formation (Yilgarn)	FR	<1840 Ma	25	45.2	40.0	1.3	2.4	26	5
Group YA dikes (Yilgarn)	YA	~1600 Ma	27	21.7	313.7	17.9	17.9	28	4
Neereno S.st. (Yilgarn)	NS	1360 ± 140 Ma	28	2.0	50.0	11.0	11.0	29	2
Morawa lavas (Yilgarn)	MLa	1360 ± 140 Ma	28	42.8	22.4	14.7	14.7	28	2
Morawa lavas (Yilgarn)	MLb	1360 ± 140 Ma	28	20.0	59.0	17.0	17.0	29	2
Campbell S.st. (Yligarn)	CS	1360 ± 140 Ma	28	31.0	75.0	16.0	16.0	29	2
Fraser dike (Yilgarn)	FD	1212 ± 10 Ma	30	55.8	325.7	4.7	5.2	33	6
Bremer Bay and Whalebone Point (Yilgarn)	BB1	~1200 Ma	31	74.4	303.8	12.2	14.7	33	6
Mt. Barren Group (Yilgarn)	MBG	~1200 Ma	31	46.6	347.4	11.9	13.9	33	6
Mundine Well dikes	MWD	755 ± 3 Ma	32	43.8	134.1	5.1	5.1	34	6
Kaapvaal craton (2.22–1.87 Ga)									
Ongeluk lava	ONG	2222 ± 13 Ma	33	–0.5	100.7	5.3	5.3	34	7
Basal Gamagara and/or Mapedi	BGM	2130 ± 92 Ma	35	2.2	81.9	7.2	11.5	35	6
Phalaborwa Complex Group 1	PB1	2060 ± 0.6 Ma	36	35.9	44.8	6.9	10.5	37	5
Lower Waterberg	WUBS1	2054 ± 4 Ma	38	36.5	51.3	10.9	10.9	39	5
Vredefort impact structure	VRED	2023 ± 11 Ma	40	22.3	40.7	11.6	15.7	35	5
Bushveld Main and Upper zones	BVMU	2061 ± 27 Ma	41	11.5	27.2	4.0	4.0	35	6
Upper Waterberg	WUBS2	1992 ± 62 Ma	39	–10.5	330.4	9.8	9.8	39	6
Hartley lava	HAR	1928 ± 4 Ma	42	12.5	332.8	16.0	16	35	4
Post-Waterberg sills	PWD	1875 ± 3.5 Ma	43	8.6	15.4	17.3	17.3	recal. from 43	5

Notes: Abbreviations: PLAT = paleopole latitude, PLONG = paleopole longitude, DP and DM = semiaxes of 95% confidence about the mean; age and paleomagnetc (Pmag) references: 1 = Odin (1994), 2 = Derder et al. (2001a), 3 = Derder et al. (2001b), 4 = Clauer et al. (1982), 5 = Teal and Kah (2005), 6 = Perrin et al. (1988), 7 = Onstott and Dorbor (1987), 8 = Onstott et al. (1984), 9 = Mortimer et al. (1988), 10 = Nomade et al. (2003), 11 = Giddings and Embleton (1976), 12 = Fanning et al. (1988), 13 = Chamalaun and Dempsey (1978), 14 = Schmidt and Embleton (1985), 15 = Wingate (1999), 16 = Embleton (1978), 17 = Arndt et al. (1991), 18 = Strik et al. (2003), 19 = Pidgeon (1984), 20 = Blake et al. (2004), 21 = Li (2000), 22 = Müller et al. (2005), 23 = Schmidt and Clark (1994), 24 = Li et al. (1993), 25 = Halilovic et al. (2004) 26 = Williams et al. (2004), 27 = Indurm (2000), 28. Giddings (1976), 29 = Indurm and Giddings (1988), 30. Wingate et al. = (2000), 31 = Pisarevsky et al. (2003), 32 = Wingate and Giddings (2000), 33 = Cornell et al. (1996), 34 = Evans et al. (1997), 35 = Evans et al. (2002), 36 = Reischmann (1995), 37. Morgan and Briden (1981), 38 = Dorland et al. (2006), 39 = de Kock et al. (2006), 40 = Kamo et al. (1996), 41 = Walraven et al. (1990), 44 = Cornell et al. (1998), 43 = Hanson et al. (2004)

contrast to the well-grouped components observed in the parent lithology (i.e., a positive conglomerate test was illustrated). Less stable magnetic components were present within two enriched BIF samples (parent lithology), and these were interpreted as being related to an intermediate-temperature magnetic overprint, named INT and ascribed to the ~1.2 Ga Namaqua orogen by Evans et al. (2002; Fig. 5B).

The stable HIG- component observed within the enriched BIF is antipodal, but otherwise similar to, the high-stability magnetic component observed in shales of the Gamagara Formation at the south Sishen mine identified during the same study at high-temperature demagnetization steps of shale samples. The pre-Gamagara age of the HIG- component, and therefore the enrichment of the BIF, is constrained by a positive conglomerate test. Evans et al. (2002), therefore, illustrated that hematitization and ore genesis were associated with oxidative weathering preceding erosion, and thus the paleomagnetic results validated suggestions that the Sishen and Beeshoek deposits formed by supergene processes during the Paleoproterozoic. By comparing the corresponding pole of the HIG- magnetic direction to other known Paleoproterozoic poles from the Kaapvaal craton (Table 1), a post-2.22 Ga but pre-2.06 Ga age can be assigned to the enrichment of the Manganore Iron Formation (Fig. 5C) and deposition of the immediately overlying Gamagara-Mapedi unit. This result was consistent with novel stratigraphic correlations of the paleosol-bearing erosional unconformity across the Kaapvaal craton (Beukes et al., 2002).

Review of Global Paleomagnetic Studies on High-Grade Hematite Deposits

Paleomagnetic studies on hard high-grade hematite deposits are few and half of them predate the use of modern paleomagnetic techniques like sequential demagnetization and principal component analysis (the latter method introduced by Kirschvink, 1980). The pre-1980 nature of much of the existing database (i.e., the possible nonremoval of secondary magnetic components and often poor statistical constraints) can explain much of the apparent lack of success of the paleomagnetic method at dating high-grade hematite ores in the past. An additional factor, in especially the older studies, is that until recently, most APWP have been poorly defined. In this contribution we reevaluate age assignments, based on updated APWP and stricter statistical criteria, and we also attempt to identify correspondence between the revised age assignments and major tectonic, erosional, and magmatic events affecting each region.

Some paleomagnetic studies have been undertaken on deposits of non-BIF–hosted hematite (e.g., Du Bois, 1962; Hodych et al., 1984), magnetite (Alva-Valdivia et al., 1990, 2000, 2003) and hematite-goethite (e.g., Kean, 1981) ore deposits. There has also been paleomagnetic work done on low-grade (<60 wt % Fe) BIF-hosted hematite (Symons, 1967c; Williams and Schmidt, 2003), BIF-hosted hematite-goethite (Symons, 1967b; Porath and Chamalaun, 1968) and BIF-hosted magnetite deposits (Symons and Stupavsky, 1979; Symons et al., 1980, 1981). Several studies have also been concerned just with the paleomagnetism of BIF or prominent BIF-bearing successions (e.g., Das et al., 1996; Schmidt and Williams, 2003; Williams et al., 2004). Such studies fall outside the scope of this paper.

Although not employing modern paleomagnetic techniques, many pioneering studies in the 1960s and 1970s were nonetheless convincingly successful at isolating hematite as the main magnetic mineral (e.g., the Marquette Range and the Vermillion Range deposits of North America, as well as the Middleback Ranges of South Australia). In doing so, the various authors were successful at eliminating less stable magnetic components that may have resided in relatively low coercivity (i.e., magnetically easily overprintable) magnetite. Two studies, respectively from the Kediat-Idjil district and the Mount Goldsworthy deposit (Gross and Strangway, 1961; Porath and Chamalaun, 1968), are excluded from this statement, because only NRM values were reported, and demagnetization was only performed on a small selection of samples to test the stability of the magnetization. Such treatment opens the possibility for the non-detection of less stable magnetic overprints. The conclusions from those two studies are considered as less reliable.

The Kediat-Idjil district, Mauritania

The first attempts at dating hard high-grade hematite deposits by the paleomagnetic method was made in the early 1960s by Gross and Strangway (1961), who sampled two limbs of the folded hard high-grade hematite orebody at the F'derik deposit in the Kediat-Idjil district. They reported only NRM values. Pilot demagnetization was performed on a selected subset of samples. These samples displayed no change in magnetic direction after demagnetization to 30 mT, but the presence of secondary magnetizations, however, cannot be ruled out because AF demagnetization was not performed on all the samples. The interpretation of Gross and Strangway (1961) that ore genesis is Precambrian in age, at least for one group of identified components (A + C1), and possibly for the B component group as well, seems justified by the paleomagnetic data. The paleomagnetic data are, however, consistent with a large range of age possibilities.

Upon inspection of the NRM values from the F'derik deposit, Gross and Strangway (1961) could identify three magnetic directions (weighted means have been recalculated from the original dataset and are presented in Figure 6, Table 2). None of the identified components are parallel to the present-day geomagnetic field. This may indicate that the hard high-grade hematite ore is unrelated to modern supergene processes. Gross and Strangway (1961) assumed younging toward the northeast and applied a structural correction to paleohorizontal accordingly. This correction causes two of the components (so-called "A" and "C1" directions) to rotate and overlap, thus alluding to a predeformational age for this magnetization. Based on this so-called "fold test" for the "A + C1" component, the F'Derik deposit was interpreted to be predominantly syngenetic, but there are several problems with this interpretation. The test is statistically inconclusive, as the precision parameter, or k value, improves merely from 6.64 to 8.32. Even if the increase in precision were significant, the structural correction would constrain the age of the A + C magnetization as being predeformation only (i.e., pre-Eburnean in age), and it would not limit the magnetization to being exclusively syngenetic.

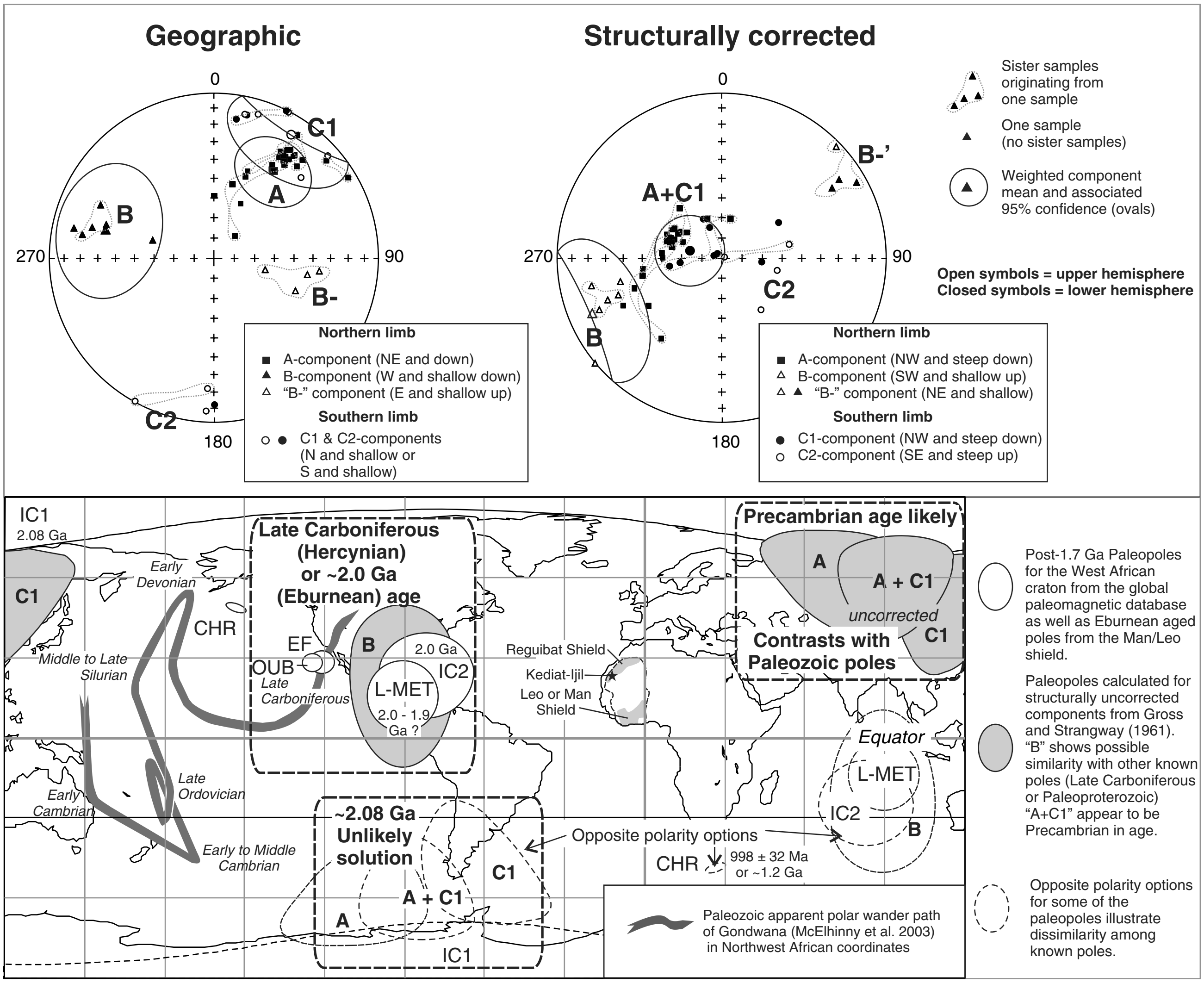

FIG. 6. Top: Equal-area plots of magnetic components, as identified by Gross and Strangway (1961), in in situ and structurally corrected coordinates. The structural correction shown assumes younging of strata toward the northeast. Values of the weighted component means are provided in Table 2. Bottom: Rectangular map projection showing virtual geomagnetic poles corresponding to magnetic components A, B, C1, and A + C1, together with poles from the global paleomagnetic database (listed in Table 1) and the Paleozoic APWP of Africa (stylized after McElhinny et al., 2003). The positions of the West African craton, the approximate position of the locality, and outlines of the continents are shown for reference only.

A second component ("B") was identified within 10 percent of the samples (all originating from the northern limb). It was interpreted by Gross and Strangway (1961) to be a secondary magnetization associated with folding and metamorphism, but no field stability tests constrain this interpretation.

The only way to assign possible ages to the identified magnetic components is to compare their corresponding virtual geomagnetic poles to other well-dated paleomagnetic poles from the Reguibat shield. Paleomagnetic data from the West-African craton are sparse and stem predominantly from the Man-Leo shield. It is uncertain exactly when the Man shield amalgamated with the Reguibat shield to form the West-African craton, but post-tectonic granites following the 2.0 to 1.8 Ga Eburnean orogeny mark the end of crustal evolution of the West African craton and it is considered stable since at least 1.7 Ga (Roussel and Lesquer, 1991). Filtered for reliability criteria, the workable database becomes a list of only three poles (Table 1, Fig. 6). Good-quality Eburnean-aged primary poles and overprints from the Man shield (an additional three poles) are also included if we assume that the Man and Reguibat shields were close to their present-day relative configuration during the Eburnean orogeny. In addition to these poles, the APWP of Africa is fairly well constrained in the Paleozoic (McElhinny et al., 2003) and comparisons can be made accordingly (Fig. 6).

Based on this limited database it can be stated that, if regarded as postdeformational, the B component bears some resemblance to Late Carboniferous poles from the West African craton (Fig. 6). Pole B, however, also shows uncanny resemblance to ~2.0 Ga poles from Liberia and the Ivory Coast (L-MET and IC2). The ~2.0 Ga poles and pole B are also similar to that of the Eburnean-aged Harper Amphibolite from Liberia (Onstott et al., 1984) and the 1950 to 1980 Ma Aftout Pluton pole of Algeria (Lomax, 1975). These poles

TABLE 2. Magnetic Components and Corresponding Pole Positions from Various High-Grade BIF-Hosted Hard Hematite Ores and Their Host Rocks

Component or sample group	n	Geographic Declin-ation	Geographic Incli-nation	Geographic $\alpha95$	Geographic k	Tilt-corrected Decli-nation	Tilt-corrected Incli-nation	Tilt-corrected $\alpha95$	Tilt-corrected k	PLAT or paleolatitude (indicated with °)	PLONG	DP limit +	DM limit –
F'Derik deposit (22.1°N, –12.7°E)													
A	6.8	34	36	18.0	12.6	–	–	–	–	58.3	74.8	12.1	20.9
B	3	285	32	29.7	18.3	247	–16	29.7	18.3	19.6	269.0	18.9	33.5
										° –7.9		7.2	–26.8
C1	3.7	32	–13	25.8	15.8	–	–	–	–	47.5	116.1	13.4	26.3
A + C1 (northward younging)	10.5	33	19	19.7	6.6	283	74	17.3	8.3	56.1	93.3	10.7	20.5
										° 59.7		90	37
A + C1 (southward younging)	10.5	33	19	19.7	6.6	323	–73	17.3	8.3	56.1	93.3	10.7	20.5
										°–58.6		–90	–36.2
Cliff-Shaft mine (46.5°N, –87.6°E)													
Negaunee Iron Fm	6		not reported			169	19	39.4	3.8	–	–	–	–
Hard ore	8		not reported			239	30	18.4	10.3	–8	216.1	11.3	20.4
										°16.1		29.4	5.9
Keweenawan (KW1)	4	213	58	50.5	4.3		not reported			–	–	–	–
Republic mine (46.5°N, –87.9°E)													
Jaspilite ore	9		not reported			212	7	93.4	1.3	–	–	–	–
Keweenawan (KW2)	4	248	72	36.0	8.1		not reported			–	–	–	–
Groveland mine (46.0°N, –87.8°E)													
Vulcan Iron Fm	4		not reported			74	41	30.6	10.0	–	–	–	–
Vermillion Range from surface, the Pioneer mine and Soudan mine (47.9°N, –92.1E)													
Ely greenstone (surface)	6	310	85	41.4	3.6		not reported			–	–	–	–
Ely greenstone (Pioneer)	3	251	65	137.4	2.0		not reported			–	–	–	–
"Paint Rock" (Pioneer)	4	90	79	30.6	10.0		not reported			–	–	–	–
Hard ore (Pioneer)	13	257	71	28.0	3.2		not reported			–	–	–	–
Soudan IF (surface)	8		not reported			338	3	16.9	11.7	39.9	117.1	8.5	16.9
Soudan IF and andesite (Soudan)	4	357	74	30.6	10.0		not reported			–	–	–	–
Altered andesite (Soudan)	4	286	54	17.1	30.0		not reported			35.0	192.9	16.8	24.0
Hard ore (Soudan)	6	182	83	19.7	12.5		not reported			34.1	267.3	37.7	38.6
Middleback Ranges (Iron Monarch = IM and Iron Prince = IP) (33.0°S, 137.0°E)													
IM + (positive inclination)	28	283	63	8.4	11.6		not reported			–14.3	91.1	10.4	13.2
IM – (negative inclination)	13	26	–58	9.9	18.6		not reported			–68.3	69.5	10.7	14.6
IP – (negative inclination)	22	63	–46	8.1	15.6		not reported			–36.1	58.9	6.6	10.4
Mount Goldsworthy (20.5°S, 119.5°E)													
Lode ore (G1)	31	317	46	5.5	22.6		not reported			19.5	84.0	6.0	6.0
Lode ore (G3)	8	115	74	8.5	45.0		not reported			30.5	329.5	12.5	12.5
Crust ore (G2)	6	263	69	12.0	31.0		not reported			21.5	259.0	19.0	19.0

TABLE 2. (*Cont.*)

		Geographic				Tilt-corrected				PLAT	PLONG	DP	DM
Component or sample group	*n*	Declin-ation	Incli-nation	$\alpha95$	k	Decli-nation	Incli-nation	$\alpha95$	k	or paleolatitude (indicated with *)		limit +	limit –
Hamersley province, at Mount Tom Price[a], Paraburdoo[b] and Mount Newman[c] (a = 22.5°S, 118.0°E; b = 23.3°S, 117.5°E; c = 23.0°S, 119.5°E)													
Mount Tom Price (TP'94)	19	309	–9	11.2	10.0	300.2	–16	14.7	6.2	–37.4	220.3	5.7	11.3
Mount Tom Price (TP'68)	28	304	25	12.0	6.1		not reported			22.0	57.0	12.0	12.0
Paraburdoo (PBD)	15	305	–23	8.3	22.2	316	–7	8.7	20.3	–36.4	209.9	4.7	8.8
Mount Newman (MN'68)	20	302	39	9.5	12.0		not reported			17.0	66.0	9.5	9.5
Hamersley province post-F2 ovpt. (HP1)	17	245	69	2.2	263		not reported			33.9	250.7	8.4	8.4
Hamersley province ?F3 ovpt. (HP2)	38	304.2	–18.1	4.5	27.6		not reported			–35.3	211.9	3.0	3.0
Beeshoek deposit (28.3°S, 23.0°E)													
Enriched Manganore Iron Fm	7	259.7	–27.6	7.7	23.2	261.4	–26.2	9.5	20.7	–0.8	276.8	10.6	20.9

Notes: Abbreviations: *n* = number of samples, $\alpha95$ = radius of 95% confidence cone about mean, k = precision parameter; underlined component means were used to calculate paleomagnetic poles, PLAT = paleopole latitude, PLONG = paleopole longitude, DP and DM = semiaxes of 95% confidence about the mean; * = paleoaltitude calculated, – = values not calculated

from the Harper Amphibolite and Aftout pluton do not pass the selection process but are summarized by Nomade et al. (2003). For simplicity they are not shown in Figure 6 or listed in Table 1.

The Man shield and the Reguibat shield were both affected by the Eburnian orogeny, but the Hercynian orogeny (300–280 Ma) affected only the northern and eastern parts of the West African craton and not the Man shield (Roussel and Lequer, 1991). Based on the paleomagnetic pole comparison the uncorrected B component may be representative either of a late Hercynian (300 Ma) remagnetization, acquired when Gondwana collided with Laurentia, or of a late Eburnian (~ 2.0 Ga) remagnetization. Since an uncorrected B direction is being compared it has to be assumed that magnetization is postdeformational.

The uncorrected A + C1 component shows no resemblance to the relatively well defined APWP of Gondwana in the Paleozoic (Fig. 6), therefore, this phase of ore genesis must have occurred in the Precambrian. The uncorrected A + C1 pole does, however, show resemblance to a ~2.1 Ga pole from the Man craton, but this option is not considered as a possibility as it relies on the comparison of post- and predeformational directions.

The uncorrected A + C1 pole is dissimilar to a pole from the I2 unit of the Char Group, for which two age options have been proposed. Clauer et al. (1982) suggested an age of 998 ± 32 Ma, whereas a more recent suggestion of ~1.2 Ga came from Teal and Kah (2005). Both of these ages are at the moment regarded as being equally unlikely for the uncorrected A + C1 magnetization due to the observed dissimilarity.

Justification of a tilt correction to any of the observed directions is difficult to evaluate. Conservatively, tilt-corrected data were treated as inclination only (i.e., they were translated into paleolatitudes and treated as such). Despite large uncertainties, the paleolatitude as defined by the tilt-corrected A + C1 magnetization bears resemblance to that defined by the 1819 to 1919 Ma steep magnetic directions reported from Aftout Gabbro and a quartz diorite from the Reguibat shield (Sabaté and Lomax, 1975). These poles, which fail the quality screening, are mentioned here but not shown in Figure 6.

The Great Lakes district, United States

The Marquette Range: The Marquette Range Supergroup of Michigan and Wisconsin is a Paleoproterozoic continental margin assemblage that lies unconformably upon the southern edge of the Superior craton. Together with several other broadly correlative successions (Ojakangas et al., 2001) the Marquette Range Supergroup forms the foreland of the Penokean orogen, a fold-and-thrust belt formed by the accretion of several island-arc terranes (collectively known as the Wisconsin magmatic terranes) to the Superior craton (Schulz and Cannon, 2007). The Penokean foreland contains the well-known Lake Superior iron formations, of which the Menominee Group of the Marquette Range Supergroup contains the largest deposits of iron ore in the region. A paleomagnetic study by Symons (1967b) targeted a hard high-grade hematite orebody, the Cliff-Shaft mine, which is hosted by the 1874 ± 6 Ma Negaunee Iron Formation (Schneider et al., 2002) of the Menominee Group. The orebody is only developed locally and stratigraphically at the very top of the iron-formation succession (Symons, 1967b). The high-grade hematite and the Negaunee Iron Formation are unconformably overlain by the Baraga Group (Schneider et al., 2002).

Twenty-seven samples were collected underground by Symons (1967b) across an easterly plunging syncline. In addition to hard high-grade hematite ore (10 samples), the

Negaunee Iron Formation host rock (8 samples) and a Keweenawan dike or sill (4 samples) were sampled. In the same year Symons (1967c) reported paleomagnetic results from oxidized BIF and Keweenawan intrusions elsewhere in the Marquette Range (Republic mine) and from BIF from two other localities in Michigan (Menominee Range and the Groveland mine). While these results were not directly from hard high-grade hematite ore they are nonetheless mentioned here, because they are significant for interpreting the results from the Cliff-Shaft mine.

Symons (1967b) employed stepwise AF demagnetization on a selection of samples and after showing that a less stable direction is removed by a 30 to 50 mT alternating field, he subjected the remainder of samples to bulk demagnetization of up to 70 mT. Eight (of the 10) hard hematite samples displayed stable remanence directions after demagnetization. Upon unfolding the syncline, the precision-parameter "k" recalculated here for the 95th percentile (Table 2) increases from 5.6 to 10.3. This suggests that the magnetization and hence ore genesis predate the deformation (a summary of the reported directions is given in Table 2). The fold test was less conclusive in dating the magnetization of the host iron formation, for which the k value decreased slightly upon unfolding (Table 2). The assumptions and conclusions, such as the fold tests of Symons (1967a, b), are difficult to evaluate, however, because he did not report pretilt declination and inclination values and means for the Negaunee Iron Formation and hard ore. Instead, only k values were reported. A wide range of ages, relative to the timing of deformation, must therefore be considered for the magnetization of the hard ore at the Cliff-Shaft mine.

Symons (1967b) regarded the remanence of the host iron formation as primary. The tilt-corrected hard high-grade hematite ore remanence differs in direction by ~70º from that of the host iron formation (Fig. 7A). Accordingly Symons (1967b) assumed a syndepositional or early diagenetic model for ore genesis to be refuted by the paleomagnetic data. However, by comparison to correlative iron-formation successions, Symons (1967c) showed that the assumed primary remanence directions from the two locations are very different (Fig. 8A), thus calling his previous assumptions into question. We choose not to draw any conclusions from the BIF results for two reasons. First, there are possible complications that may arise from anisotropy of magnetic susceptibility (e.g., Clark and Schmidt, 1993) and second, the high uncertainty in the results (Table 2) makes interpretation very difficult.

Concerning the age of folding at the Cliffshaft mine, development of the Penokean fold-and-thrust belt (Fig. 7A) is

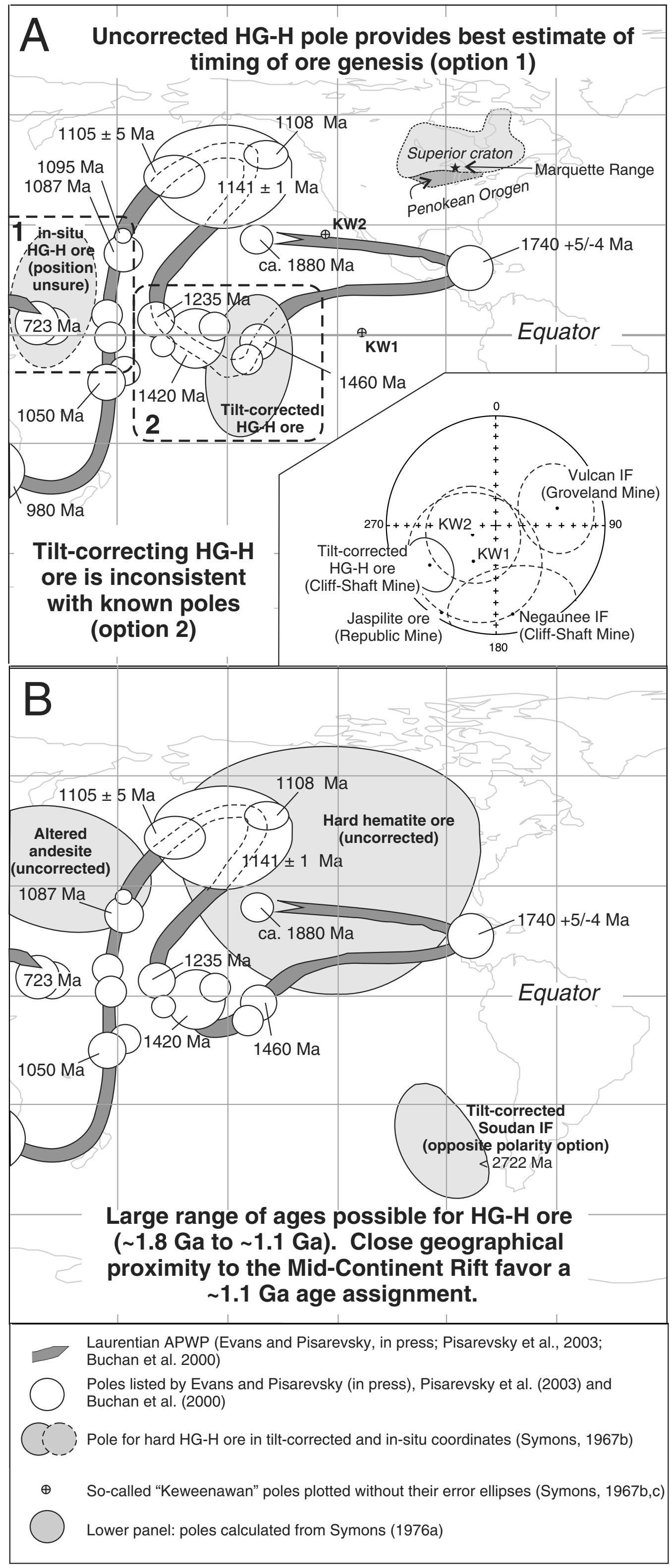

FIG. 7. A. Rectangular map projection showing a virtual geomagnetic pole obtained for hard hematite ore from the Cliff-Shaft mine, as well as poorly constrained virtual geomagnetic poles for the so-called "Keweenawan" intrusions. The APWP for Laurentia during the Mesoproterozoic to early Neoproterozoic (Buchan et al., 2000; Pisarevsky et al. 2003; Evans and Pisarevsky, in press) is shown for comparison. The APWP of Laurentia during the Phanerozoic differs significantly from the poles obtained from the Cliff-Shaft mine and is not shown in this figure. The outlines of the Superior craton, Penokean orogen, approximate sampling locality and continents are shown for reference only. The inset at the bottom right shows mean magnetic components obtained by Symons (1967b, c) in geographic coordinates. Closed symbols = lower hemisphere, ovals = confidence limits and are stippled when larger than 15º. B. Rectangular map projection showing virtual geographic poles obtained from hard hematite ore and for highly altered, hematite-impregnated andesite as well as iron formation host rock (Soudan Iron Formation). The APWP of Laurentia during the Mesoproterozoic and early Neoproterozoic is shown for comparison.

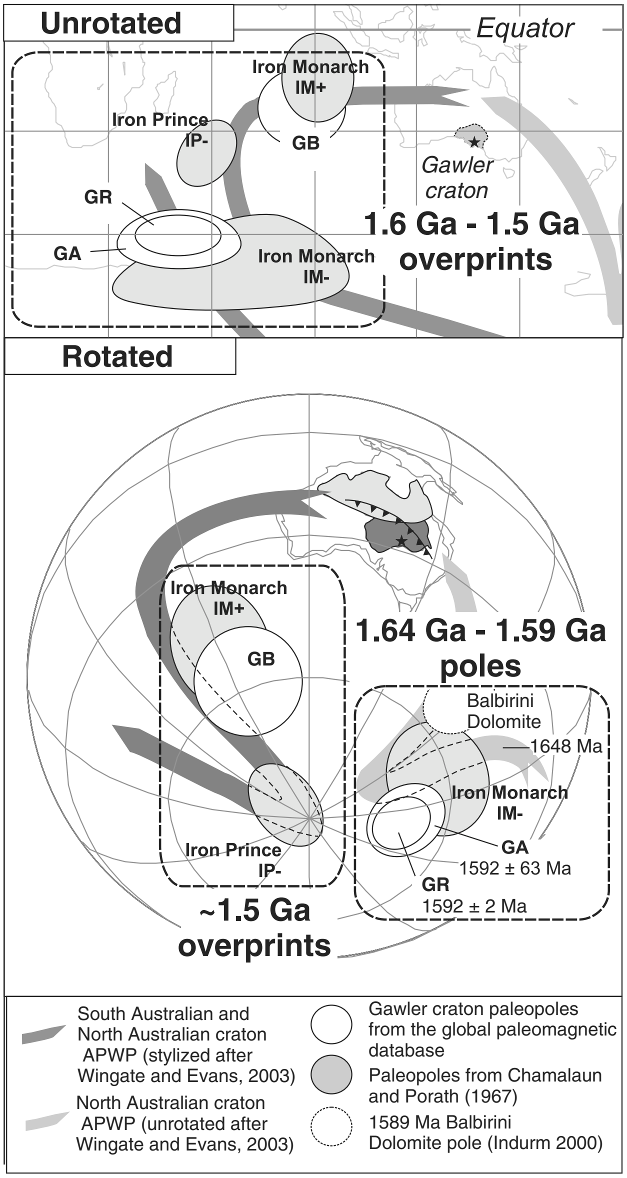

FIG. 8. Top: Rectangular map projection showing virtual geomagnetic poles for components identified by Chamalaun and Porath (1968), as well as known poles from the Gawler craton (listed in Table 1). The APWP of the South and North Australian cratons between 1600 and 1500 Ma (stylized after Wingate and Evans, 2003) are shown for comparison. The outline of the Gawler craton, approximate sampling locality, and continents are shown for reference only. Bottom: An alternative reconstruction of the Australian cratons is shown, in which the Gawler craton is rotated relative to the North Australian craton (Giles and Betts, 2000; Wingate and Evans, 2003). Different age interpretations for ore genesis are made possible with this reconstruction.

thought to have occurred during a late stage of the Penokean orogeny between ~1850 and ~1830 Ma (Schulz and Cannon, 2007). The authors of that review, however, pointed out that a younger tectonic event, correlated to the ~1.63 Ga Mazatzal orogeny, affected parts of northern Michigan. The folding of both the Penokean and Mazatzal deformations are approximately coaxial, and according to Schulz and Cannon (2007) it is possible that some fold structures previously ascribed to Penokean deformation might in fact be Mazatzal in age. The ore genesis, if predating the deformation as Symons (1967b) has concluded, may therefore be either older than 1.85 to 1.83 Ga or it may be merely older than 1.63 Ga.

The APWP of Laurentia (Fig. 7A) is fairly well constrained for the time interval following the deposition of the Negaunee Iron Formation, particularly in the interval between 1460 and 1080 Ma (Buchan et al., 2000). The APWP before this time is suggested by the ca. 1880 Ma Molson dikes pole and the 1740^{+5}_{-4} Ma Cleaver dikes pole to be represented by a swath from a position off the coast of Baja California moving first east-southeast and then southwest toward the 1460 Ma position (Evans and Pisarevsky, in press). In addition, Pisarevsky et al. (2003) review the so-called Grenville loop and younger APW motion between 1080 and 723 Ma. The ore pole can also be compared with the Phanerozoic APWP of Laurentia (e.g., Van der Voo, 1993).

The structurally corrected pole for the hard hematite ore at the Cliff-Shaft mine is different from directions expected for Penokean and Mazatzal times, and it does not bear any resemblance to the APWP of Laurentia in the Phanerozoic (out of the plotted area in Fig. 7A). It does, however, bear resemblance to the 1460 to 1200 Ma poles of Mesoproterozoic APWP of Laurentia (Fig. 7A). Symons (1967b) assumed a predeformational age for the magnetization (i.e., pre-Penokean or pre-Mazatzal, or pre-1.85–1.83 and pre-1.63 Ga, respectively). This is in stark contrast with the Mesoproterozoic ages given by the APWP method and implies that either the data are not to be trusted or application of the structural correction is incorrect (our favored explanation). Unfortunately, no in situ data were reported by Symons (1967b), but from the locality map in their paper a general west-southwesterly strike and dip of ~50º can be assumed for the hard ore. This information can be used to back orrect the hard ore mean. Such back-correction results in an approximated declination of ~270º and inclination of ~25º, which translates to a virtual geomagnetic pole that is comparable with the APWP segments between either 1095 and 1087 or 780 and 723 Ma (Fig. 7A). Both these age options are considered equally plausible.

A defendable case can thus be made for a postdeformational late Mesoproterozoic to middle Neoproterozoic age for the magnetization identified by Symons (1967b) in hard hematite ores from the Marquette Range. A predeformational age option appears to be contradicted by later paleomagnetic studies defining the APWP with better reliability.

The Vermillion Range: Symons (1967a) sampled hard ore from deposits in the Soudan Iron Formation in the ~2.7 Ga Ely greenstone belt (Peterson et al., 2001). In addition, samples were also taken of the iron-formation host rocks and interbedded, significantly altered, andesite. Alteration includes chloritization and impregnation by hematite (Symons, 1967a). Samples originated from both surface outcrops and from underground workings (Pioneer and Soudan mines). The demagnetization procedure was identical to that of Symons (1967b) and all precision parameters and confidence limits were reported at the 63 percent confidence level. These values were recalculated for the 95^{th} level (Table 2). Use of the dataset is hampered by the fact that either geographic or tilt-corrected means, but never both, were reported. Symons (1967d) cited a significant decrease in precision upon structural correction for the altered andesite (k = 9.8–1.4) and ore means (k = 29.5–17.3) as an argument for a postfolding age for the ore formation.

The corresponding virtual geomagnetic poles for the uncorrected means, in particular that of the altered andesite, show similarities to the ~1.1 Ga Keweenawan paleopoles (Fig. 7B) of Laurentia (as listed by Buchan et al., 2000). The apparent postfolding age of magnetization and similarity to the Keweenawan-aged thermal event, as well as the close geographical proximity of the study area to the Mid-Continent rift, is suggestive of ore genesis at ~1.1 Ga or a remobilization of iron at ~1.1 Ga. It is noted, however, that the large uncertainty in the high-grade hematite ore pole makes possible a range of age assignments (~1.8–~1.1 Ga; Fig. 7B).

The Middleback Ranges, South Australia

The Middleback Ranges, a discontinuous series of north-south–oriented hills in the Eyre peninsula of South Australia, produced the first high-grade iron ore to be mined in Australia (Yeats, 1990). The Hutchison Group, of which the Middleback Subgroup forms part, overlies the Archean Sleaford Complex basement (Gawler craton). The whole area has undergone a complex history of intrusion by granites and basic intrusions, culminating in the ca. 1.73 to 1.70 Ga Kimban orogeny (Wingate and Evans, 2003; Swain et al., 2005).

The western part of the Gawler craton is affected by the poorly understood ca. 1650 to 1540 Ma Kararan orogeny (Swain et al., 2005). This event culminated with the emplacement of ca. 1620 Ma (Flint et al., 1990) tonalitic and granodioritic rocks of the St. Peter Suite in the central Gawler craton.

The Hutchison Group is unconformably overlain by the ~1592 Ma felsic volcanic rocks of the Grawler Range Volcanics and associated Mesoproterozoic clastic sedimentary rocks (Drexel et al., 1993), including the Corunna Conglomerate. The presence of enriched hematite pebbles within the Corunna Conglomerate (Yeats, 1990) constrains ore formation to before 1592 Ma.

According to Yeats (1990), orebodies preferentially occur in synclinal keels of the Middleback Subgroup where iron formation has been structurally thickened. Furthermore, basic intrusions associated with the final stages of the Kimban orogeny appear to have acted as barriers for ore-forming fluids (Yeats, 1990). Orebodies are also cut by 825 Ma northwest-southeast–trending dolerite dikes (Wingate et al., 1998) of the Gairdner Dike Swarm (Chamalaun and Porath, 1968; Yeats, 1990).

Chamalaun and Porath (1968) undertook a systematic study of the most accessible orebodies available at the time. These were the Iron Prince, Iron Monarch, and Iron Duke deposits (Fig. 8). The Iron Duke deposit unfortunately did not yield any reliable results. Ore samples were in most cases subjected to thermal demagnetization to 600°C or AF demagnetization to 75 mT (Chamalaun and Porath, 1968). This treatment removed small components of secondary magnetization and effectively isolated hematite as the main carrier of remanence. Two stable groups of magnetization of opposite polarity (nonantipodal) were revealed. Both groups are well developed in the Iron Monarch deposit, while negative inclination magnetizations are generally recorded by samples from the Iron Prince deposit (Chamalaun and Porath, 1968). These directions correspond to virtual geomagnetic pole positions (Table 2, Fig. 8) that are far removed from what is expected of the present-day geomagnetic field. Contrasts with Phanerozoic paleomagnetic poles from Australia (Chamalaun and Porath, 1968) and Africa (Porath, 1967) led these authors to assign a Precambrian age to both magnetization groups. These analyses, however, were severely hindered by the scarcity of Precambrian paleomagnetic poles. In subsequent years more poles have become available from the Gawler craton (Giddings and Embleton, 1976; Chamalaun and Dempsey, 1978) and a 1.6 to 1.5 Ga age (Chamalaun and Dempsey, 1978; Wingate and Evans, 2003) was assigned to the deposits at Iron Monarch and Iron Prince (Fig. 8, top panel). In this interpretation the poles from the 1592 Ma Gawler Range Volcanics as well as the poles from the Iron Monarch and Iron Prince hard high-grade hematite ores are believed to be later overprints. Note that this age assignment assumes that the South Australian craton (including the Gawler craton) and North Australian craton moved together as a unit between 1600 and 1500 Ma. Another proposed reconstruction, in which the South Australian craton is rotated anticlockwise 55° with some eastward translation relative to the North Australian craton (Giles and Betts, 2000; Wingate and Evans, 2003), not only better aligns linear Paleoproterozoic-Mesoproterozoic elements but also better aligns the Gawler Range Volcanics pole with a coeval pole from Northern Australia (Balbirini Dolomite; Fig. 8, bottom panel). Application of the Giles and Betts (2000) rotation therefore suggests that the Gawler Range Volcanics pole is primary. A similar age can then be assigned to the Iron Monarch negative inclination directions, based upon its similarity to the Gawler Range Volcanics and Balbirini Dolomite poles (Fig. 8, bottom panel). The other Iron Monarch pole (IM+) and the Iron Prince pole (IP-) then appear to be younger overprint directions that developed around 1.5 Ga (Fig. 8, bottom panel).

The Giles and Betts (2000) rotation, however, does not bring the Gawler Range Volcanics and Balbirini Dolomite poles in perfect alignment, and Wingate and Evans (2003) suggested that the model of Giles and Betts (2000) might need to be modified. Another possibility is that either of the Gawler Range Volcanics or Balbirini Dolomite poles could be representative of later overprints. The paleomagnetic database at present, however, is inadequate at distinguishing between the single stable continent model at ~1.6 Ga versus minor rearrangement of constituent cratons as proposed by Giles and Betts (2000).

The paleomagnetic data suggest multistage ore genesis and are currently consistent with several age options near the end of the Paleoproterozoic and early Mesoproterozoic eras. In one option, ore formation is associated with the uplift that preceded deposition of the Corunna Conglomerate and might possibly be related to intrusion of the St. Peter Suite at the end of the Kararan orogeny or with the volcanism of Gawler Range Volcanics. In this option initial formation of ore is followed by two later stages of ore genesis at ca. 1.5 Ga. Another interpretation favors ore genesis taking place as a series of events between 1.6 and 1.5 Ga after the deposition of the Corunna Conglomerate and extrusion of the Gawler Range Volcanics. The first option, which involves rotation of the Gawler craton, is more consistent with geologic constraints (presence of ore pebbles in the Corunna Conglomerate), but the second interpretation cannot be excluded because the paleomagnetic directions may be representative of later chemical remagnetization. A conglomerate test on the Corunna ore pebbles could distinguish between these options.

Mount Goldsworthy, Western Australia

A systematic study of hematite orebodies from Western Australia by Porath and Chamalaun (1968) included deposits of the Hamersley province and the Archean-hosted Mount Goldsworthy deposit in the Pilbara craton. Their study also included the Koolyanobbing deposit of the Yilgarn craton, a mixed iron oxide and predominantly goethitic deposit that falls outside the scope of this paper.

The Mount Goldsworthy deposit is developed in BIF of the 3.24 to 2.94 Ga (Van Kranendonk et al., 2002) Gorge Creek Group and is, with 65 Mt of premining reserves, the largest iron ore deposit in the north Pilbara granite-greenstone terrane (Geological Survey of Western Australia, 1990).

Natural remanent magnetization directions were reported (Table 2) from iron formation host-rock, hard high-grade hematite ore or so-called "lode ore," as well as from hematite surface crust. According to Brandt (1966) the crust or surface ore (hematite-goethite) is developed at upturned edges of steeply dipping iron formation within the Tertiary duricrust. The lode ores are expressed as bedding-parallel lenses of massive hematite and are associated with normal faults cutting the steeply dipping BIF (Brandt, 1966). Hematite ore pebbles are found in the overlying Phanerozoic conglomerate (Brandt, 1964).

Thermal demagnetization and AF demagnetization to 75 mT on presumably representative samples indicated a single Curie temperature at 695ºC and high stability against AF demagnetization (Porath and Chamalaun, 1968). Unlike most other early paleomagnetic studies where bulk demagnetization on all samples was performed to temperatures in excess of 600ºC (e.g., Chamalaun and Porath, 1968), the presence of secondary magnetic components cannot be ruled out for samples from the Mount Goldsworthy deposit. Porath and Chamalaun (1968) did not make or report any structural corrections for the dip of the strata, thereby making it more difficult to evaluate the age of magnetization. Two magnetic directions (named G1 and G3) were nevertheless reported from the lode ores at Mount Goldsworthy, whereas one well-grouped mean (G2) was determined for the crust ore. The directions are regarded with caution.

The G1 and G3 directions are structurally uncorrected and a postdeformational age for the magnetization was assumed by Porath and Chamalaun (1968). The east Pilbara granite-greenstone terrane (including the Gorge Creek Group) was deformed at ~2935 Ma before the final cratonization of the north Pilbara block at 2.85 Ga (Van Kranendonk et al., 2002). When compared to known poles from the Pilbara craton, between 2.95 Ga and 542 Ma, and post-1.8 Ga poles from the Yilgarn craton, which was amalgamated as the West Australian craton from this time onward (Tyler and Thorne, 1990; Evans et al., 2003), several age possibilities for remanence directions G1 and G3 are revealed (Table 1, Fig. 9). The G3 remanence (positive inclination) can be assigned a ~2.77 or ~1.6 Ga age. Note that the APWP for the Pilbara craton between 2.8 and 2.7 Ga (Strik et al., 2003) is far better constrained than that for the West Australian craton between 1.7 and 1.6 Ga, by which time it is possible that the Australian craton was assembled (Idnurm, 2000; Wingate and Evans, 2003). The opposite-polarity option of the G3 pole shows no resemblance to any of the known poles. The opposite polarity G1 pole bears some similarity to the preliminary pole for the ~1.36 Ga Morawa Lavas of the Yilgarn craton and plots near the 1.36 to 1.32 Ga segment of the Australian APWP (Idnurm and Giddings, 1988; Wingate and Evans, 2003). The paleomagnetic data from the Mount Goldsworthy hard high-grade hematite ore suggest an Early or Late Paleoproterozoic age for some of the hematite (G3) and a Middle Mesoproterozoic age for the remainder (G1), but the quality of the data is questionable.

The questionable quality of the data becomes apparent when looking at the G2 crust-ore pole, which is widely thought to have developed during the Tertiary. The G2 pole, however, does not bear any resemblance to Tertiary or late Mesozoic poles from Australia, indicating that the observed magnetic direction (G2) is not representative of the ore-forming event during the Phanerozoic. It is regarded, therefore, as either some spurious magnetization or some unidentified overprint direction.

The Hamersley province, Western Australia

The Hamersley province of the Pilbara craton contains some of the world's largest iron ore deposits. The deposits are hosted within the ca. 2600 Ma Marra Mamba Iron Formation and the ca. 2490 to 2460 Ma Brockman Iron Formation of the Hamersley Group (cited in Nelson et al., 1999) and have received the most attention in terms of paleomagnetic studies of ore deposits of this kind (Porath, 1967; Porath and Chamalaun, 1968; Clark and Schmidt, 1993; Li et al., 1993; Schmidt and Clark, 1994). The age of ore formation is constrained by the first appearance of enriched ore pebbles in conglomerates that overlie the BIF (Morris, 1985, among others). The conglomerates of the 2031 ± 6 Ma (Müller et al., 2005) Lower Wyloo Group are barren in this respect, whereas ore pebbles exist within the basal conglomerate of the Mount McGrath Formation of the Upper Wyloo Group. An independent age constraint for the ore formation comes from its relationship with mafic intrusions. Pre-Lower Wyloo Group dikes, which were folded during the Ophthalmian orogeny, were recently dated by Müller et al. (2005) at ca. 2208 Ma. Another set of undeformed intrusions (northwest-trending) has a close genetic relationship with the iron ores. They cut the Lower Wyloo Group, but do not intrude the Mount McGrath Formation and have recently been dated at ca. 2008 Ma (Müller et al., 2005). These new ages constrain the ore formation as postdating the Ophthalmian orogeny and as being closely related to the 2008 Ma mafic dikes that intruded the Lower Wyloo Group.

Following some early studies in the late 1960s (Porath, 1967; Porath and Chamalaun, 1968), which suggested a Precambrian age for the orebodies, Schmidt and Clark (1994) resampled hard high-grade hematite ore from Mount Tom Price and Paraburdoo (see also Clark and Schmidt, 1993) and reported on unpublished results from Mount Newman. Those authors also sampled oxidized BIF from the two deposits, thereby expanding on the paleomagnetic work done by

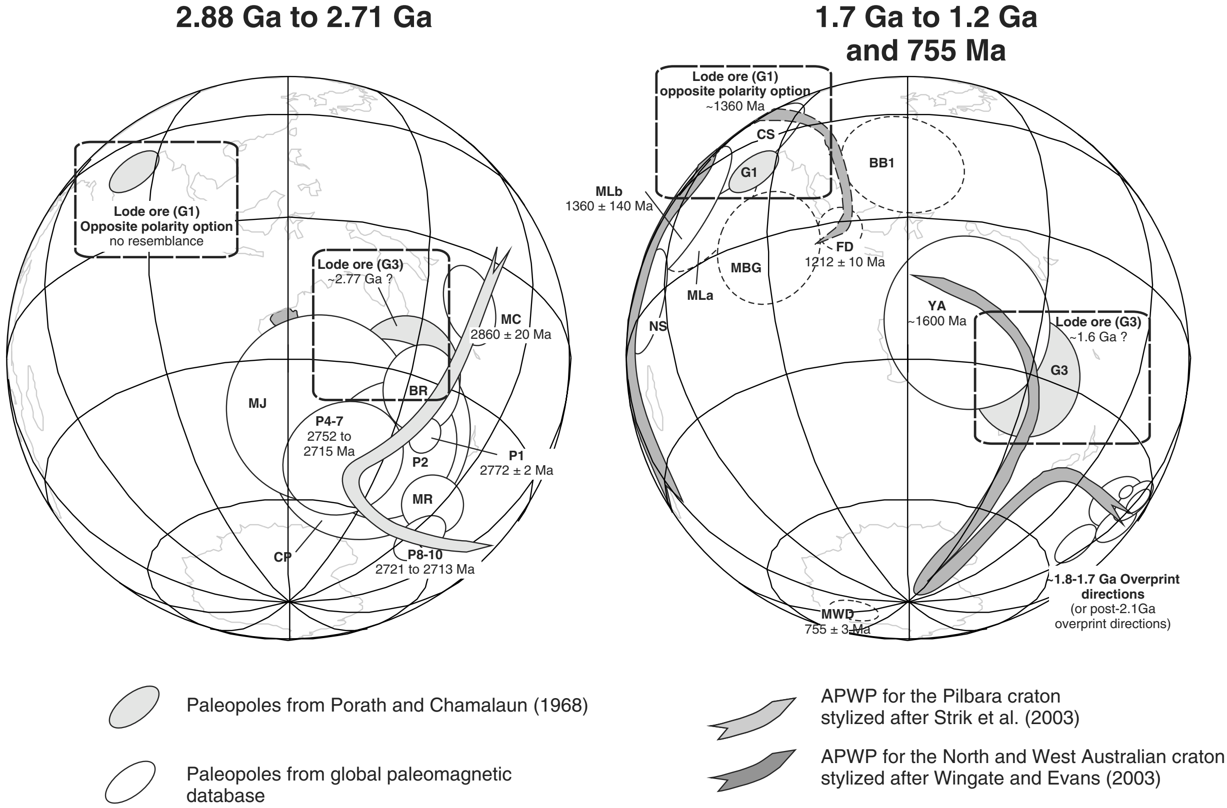

FIG. 9. A view from space of virtual geomagnetic poles obtained from lode ore at the Mount Goldsworthy deposit (Porath and Chamalaum, 1968) and known poles for the Pilbara craton (listed in Table 1) are shown together with known poles younger than 1.8 Ga from the Yilgarn craton (listed in Table 1) and suggested APWP for the Pilbara craton in the late Neoarchean (stylized after Strik et al., 2003) and the west Australian craton in the Paleoproterozoic to Mesoproterozoic (stylized after Wingate and Evans, 2003).

Embleton et al. (1979) on the BIF host rocks. By making a correction for the anisotropy of the BIF, Schmidt and Clark (1994) illustrated a positive fold test for the magnetization carried by the BIF. They suggested the acquisition of this magnetization to have taken place shortly before the Opththalmian orogeny, based upon the similarity of the magnetization of the BIF and that of the synfolding Mount Jope Formation overprint (Schmidt and Embleton, 1985). The corresponding poles to the high-grade hematite ore were indistinguishable from the poles obtained from the BIF (Fig. 10) and indistinguishable from the Mount Jope Formation overprint (Schmidt and Clark, 1994). Schmidt and Clark (1994) suggested an early post-Ophthalmian age for the ore formation and suggested supergene enrichment during uplift and exposure as the ore-forming process (Fig. 10, top right). In opposition to this view, Li et al. (1993) illustrated the presence of similar magnetizations within other lithologic units (e.g., shale and dolostone) and units (e.g., Fortescue Group and Jeerinah Formation) of the Mount Bruce Supergroup (Table 1, Fig. 10) and later referred to similar magnetic directions in rocks as young as 1.7 Ga (Li, 2000). Li et al. (1993) regarded the positive fold test of Schmidt and Clark (1994) and Clark and Schmidt (1993) as representative of the Capricorn orogeny and instead suggested that the magnetic directions are representative of a pervasive magnetic overprint that affected the Hamersley basin at ca. 1.8 to 1.7 Ga (Fig. 10, center right). The ambiguity concerning the age of the Pilbara craton overprint directions exists because of different interpretations of the age of folding. Both Ophthalmian and Capricorn ages have been applied to the remanence directions recorded by the hematite ore.

Accepting the geochronological constraints on ore formation (Müller et al., 2005), it seems a likely option that iron ores developed shortly after the Ophthalmian orogeny and that their remanence direction is mimicked by or overprinted by the younger 1.7 Ga remanence that is observed regionally within the Hamersley province. This interpretation is consistent with 1.8 to 1.7 Ga poles from the North Australian craton as well as a Yilgarn craton pole from the Frere Formation (Williams et al., 2004) of the <1.84 Ga Earaheedy basin (Halilovic et al., 2004; Pirajno et al., 2004).

Within an unpublished report, Li et al. (2000) added another complexity that must be considered when evaluating the timing of Hamersley ore genesis by the paleomagnetic method. Apart from observing the prominent presumably 1.8 to 1.7 Ga overprint direction within hard high-grade hematite ore and other rock units from the Hamersley province (collectively named HP2), Li et al. (2000) also reported a steep downward magnetic direction (named HP1) from high-grade hematite ores from the eastern parts of the Hamersley province (Fig. 10). A negative fold test illustrates the HP1 direction to be post-Ophthalmian in age (Li et al., 2000), but

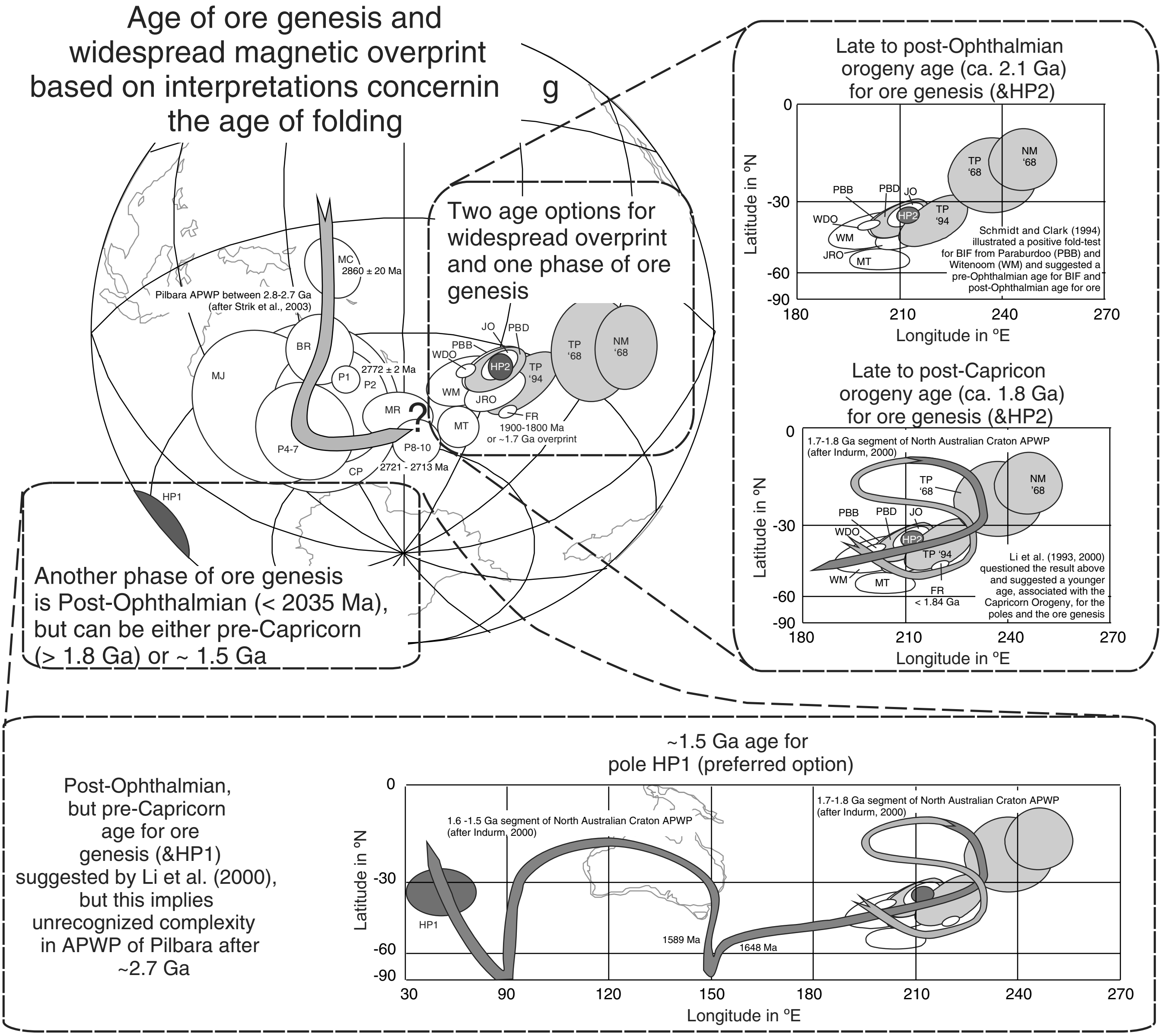

FIG. 10. Equal-area projection of virtual geomagnetic poles from hard hematite ore of the Hamersley province, together with known late Neoarchean to Paleoproterozoic poles from the Pilbara craton (listed in Table 1). The APWP for the Pilbara craton as suggested by Strik et al. (2003) is shown for comparison. Shown to the right are two possible age interpretations for the ore formation and development of a significant magnetic overprint component within rocks of the Hamersley province. The bottom panel address the presence of multiple phases of ore genesis (reported by Li et al., 2000). Two age assignments are possible for the so-called HP1 pole. Segments of the APWP of the North Australian craton (stylized after Indurm, 2000; Wingate and Evans, 2003) are used to illustrate the different age possibilities.

depending on how structures are interpreted, the direction may be younger or older than the 1.8 to 1.7 Ga Capricorn orogeny. If the direction were pre-Capricorn, as Li et al. (2000) suggested, it would imply some unrecognized complexity to the APWP of the Pilbara craton after 2.7 Ga in order to accommodate the position of the HP1 pole (Fig. 10, bottom left). Li's coauthors were of the opinion that the HP1 direction postdates the Capricorn orogeny (Li et al., 2000), and when it is compared with the APWP of the north Australian craton (Idnurm, 2000) an age of ~1.5 Ga seems likely (Fig. 10, bottom right). Note that this comparison assumes that the north Australian craton has been completely amalgamated with the west Australian craton since 1.5 Ga (see discussion in Wingate and Evans, 2003). The higher stability of the HP1 component (compared to that of the HP2 component when both components were observed together in the same sample) has been interpreted by Li et al. (2000) to imply that the HP1 remanence is older. It is noted here, however, that if the HP1 direction is indeed a chemical remanence as suggested by Li et al. (2000) then the order in which components unblock during thermal demagnetization is unrelated to timing of remanence acquisition.

In summary, it can be said that the paleomagnetic data are definitely supportive of a post-Ophthalmian age for ore genesis. Based on age interpretations of folding, ore genesis may have occurred either shortly after the Ophthalmian orogeny at ~ 2.1 Ga or shortly after the Capricorn orogeny at ~ 1.8 Ga. Multiple magnetic components within high-grade hematite ore (HP1 and HP2 of Li et al., 2000) must be considered. Uncertainty in the APWP of the Pilbara craton between 2.7 and 1.8 Ga makes it possible to say that the HP1 direction was acquired shortly after ~2.1 Ga (Li et al., 2000) but doing so would imply that the HP2 direction and other overprint directions are not close to ~2.1 Ga in age, but rather 1.8 to 1.7 Ga old. Another possibility may be that the HP2 direction was acquired first, at ~2.1, 1.8 to 1.7 Ga, or possibly both times. In this case the HP1 direction could be a younger overprint at ~1.5 Ga. In any of these options, the multiple phases of ore formation are in accordance with the protracted span of hydrothermal xenotime U-Pb ages of Rasmussen et al. (2007); however, the predominance of one or two paleomagnetic directions implies that the ores were formed in discrete episodes within that broad time span.

Distribution of Hard High-Grade Hematite Deposits through Time

Although the conclusions from the Kediat-Idjil and Mount Goldsworthy deposits are considered less reliable and allow for a broad range of age possibilities (Fig. 11), paleomagnetic data do seem to suggest a Precambrian age as one possibility for the Kediat-Idjil deposit, and independent geologic evidence (Brandt, 1964) constrains the Mount Goldsworthy deposit as being Precambrian in age. The Precambrian age of these deposits is something they appear to have in common with the other evaluated hard high-grade hematite deposits (Fig. 11).

Early studies (Porath, 1967) pointed toward a Precambrian origin for the Hamersley orebodies of the Pilbara craton of Western Australia. Precambrian ore genesis of the Hamersley deposits is supported and further constrained to the Paleoproterozoic and Mesoproterozoic by more recent paleomagnetic studies (Clark and Schmidt, 1993; Li et al., 1993, 2000; Schmidt and Clark, 1994) and mineralogical-geochronological work (Müller et al., 2005).

Early paleomagnetic studies of ore deposits from the Great Lakes district have suggested a Precambrian age for ore formation (Symons, 1967b, a), but our reevaluation is more precise, indicating a Meso- or Neoproterozoic origin of the deposits in the Marquette Range and a Mesoproterozoic origin for the Vermillion Range deposits.

Paleomagnetism has also validated a Paleoproterozoic age for the Sishen and Beeshoek deposits of the South Africa (Evans et al., 2002). Elsewhere on the Kaapvaal craton, an age of about 2.04 Ga has been proposed for the Thabazimbi deposit (this study).

The relatively old age of the deposits and conspicuous lack of Phanerozoic hard high-grade hematite ore is unexpected at first glance, since many proposed models rely upon processes that were active since the Precambrian and throughout the Phanerozoic up to the present day. It is acknowledged that the exclusive Precambrian ages may be the result of a limited database, as may be evidenced by the Kediat-Idjil deposit, which shows possible evidence of an ore genesis episode or iron remobilization as young as ~300 Ma.

Relationships based on paleomagnetically estimated ages, between ore genesis and major tectonic, erosional, and magmatic events are not simple (Fig. 11). The Kediat-Idjil deposit displays one phase of ore genesis that either just postdates the late Carboniferous Hercynian orogeny or just postdates the ca. 2.0 Eburnean orogeny. The Marquette Range deposits, in contrast, display no overlap with times of potential BIF deformation. Although it is a possibility that one phase of ore genesis in the Vermillion Range deposit just postdates the Penokean orogeny, it is far more likely that ore genesis took place much later. Both the Marquette Range and Vermillion Range deposits display an apparent association with widespread magmatism at ~1.1 Ga. The Middleback Ranges of Southern Australia is another example of hard high-grade hematite ore deposits that have an association with a magmatic event (i.e., with the Gawler Range Volcanics) rather than a deformational event (the Kimban orogeny). It is worth mentioning that there is a possibility that one phase of ore genesis could have been associated with the so-called Kararan orogeny. In order for this association to be valid, the Gawler craton must be rotated relative to the North Australian craton. For other deposits, like the Mount Goldsworthy deposit of Western Australia, the quality of the paleomagnetic data makes it very difficult to come to any defendable conclusions about the timing of ore genesis. Elsewhere on the Pilbara craton of Western Australia, deposits of the Hamersley province again show strong associations with deformation events, and deposits can be related to either the Ophthalmian or the Capricorn orogeny, or both. A third possible phase of ore genesis in the Hamersley province may have occurred at around 1.5 Ga and does not have any obvious associations with periods of uplift, deformation, magmatism, or metamorphism.

The relative ages of the South-African hard high-grade hematite deposits are probably the best constrained of any in the world. Although no direct age data are available from the basal Gamagara-Mapedi unit, paleomagnetism does indicate that the Sishen-Beeshoek deposits are intimately associated with the development of the unconformity at the base of that unit (Evans et al., 2002). The host BIF of the so-called Manganore Iron Formation experienced long-lived exposure during which the unconformity at the base of the Gamagara-Mapedi unit developed (Evans et al., 2002). The Sishen-Beeshoek deposits are not associated with a short-lived deformational event but rather with the development of karst depressions and sinkholes within the underlying Malmani dolostones. It is to these sinkholes that the Sishen and Beeshoek deposits owe their preservation. New data from the Thabazimbi deposit indicate ore genesis to postdate closely the deformation related to thermal relaxation of the crust following intrusion of the Bushveld Complex. Although the deposit is closely associated in time and space to the Bushveld Complex, no direct link between this large magmatic event and the iron ore is yet apparent.

The paleomagnetic results from the two South African examples discussed in this paper show very good agreement with what is predicted by proposed models for ore genesis at these deposits. Additionally, paleomagnetic data suggest previously unconsidered phases or times of hematite development, for

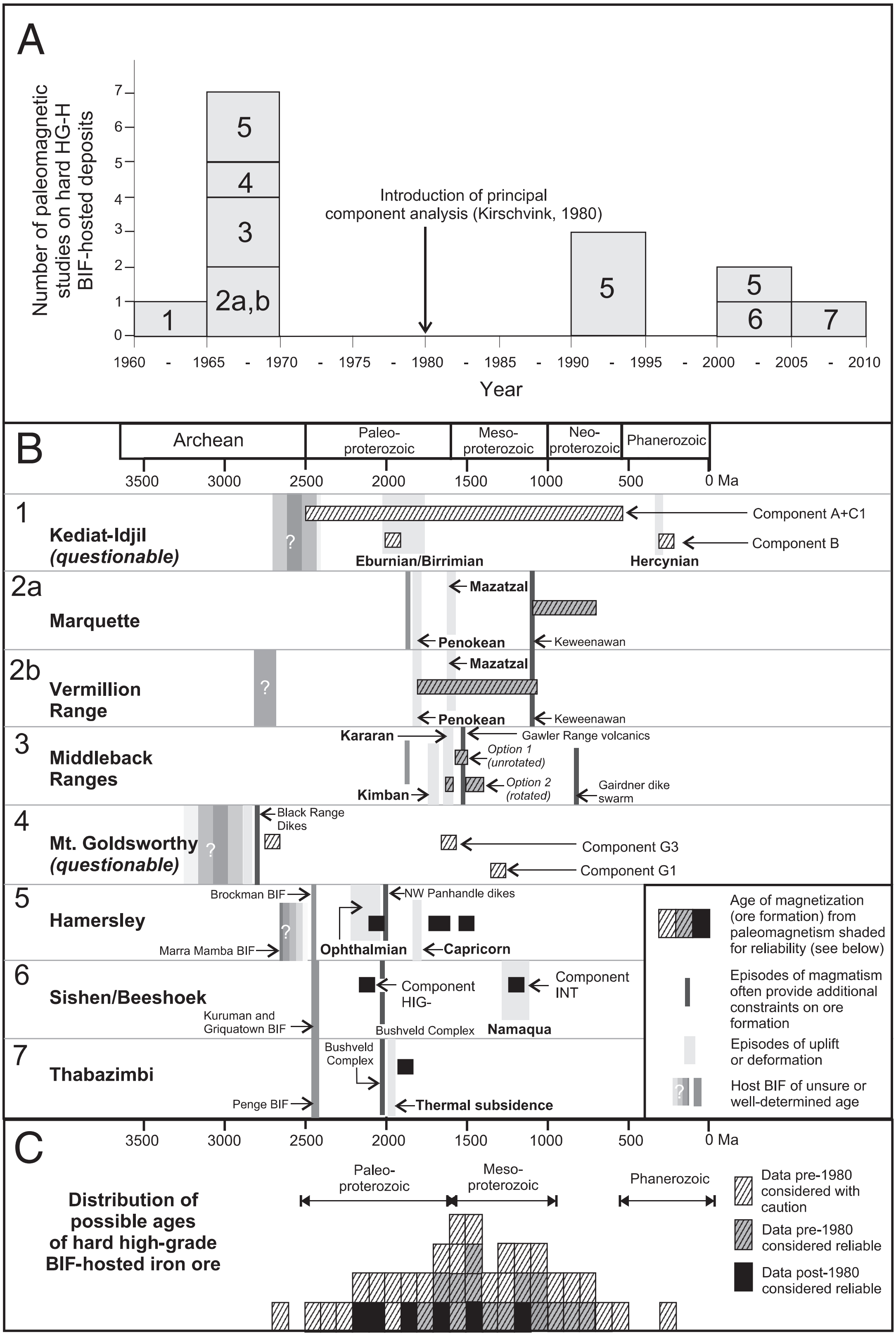

FIG. 11. A. Time distribution of paleomagnetic studies on high-grade hematite deposits. Numbers refer to deposits as listed in (B). B. Ore genesis of high-grade BIF-hosted hard hematite ores through time and their possible relationship to geotectonic, erosional, and magmatic events. C. A spectrum of possible ages for these deposits (as indicated by the paleomagnetic method). Data acquired after 1980 is considered more reliable than earlier studies and some of the pre-1980 studies are considered more reliable than others.

example, the development of minor neoformed hematite in the Sishen-Beeshoek deposits during the Namaqua orogeny, as illustrated by the presence of the so-called INT magnetic component. By extension, the paleomagnetic data assembled for other deposits may be revealing of their respective origins, for example, single or multiple phases of ore formation associated with major magmatic events (e.g., possibly in the cases of the Great Lakes district and the Middleback Ranges). Care should be taken, however, since proposing ore-forming events via paleomagnetic directional comparisons, without any independent geochronological constraints, may prove erroneous. The Thabazimbi deposit's close association with, but isolated development from, the Bushveld complex serves as a good example.

Conclusions, Exploration Significance, and Future work

The paleomagnetic method is particularly well suited for testing proposed ore genesis models and age estimates of hard high-grade hematite deposits hosted by BIF, as illustrated by two examples from South Africa. The usefulness of older (pre-1980) studies is often limited by outdated demagnetization methods, and even more recent efforts can be hampered by less well-constrained apparent polar wander paths. The latter factor imposes the greatest limitations on the method at present. Regular reevaluation of age assignments is therefore recommended in the future, as updated APWP become available.

The present evaluation of existing studies indicates that most hard high-grade hematite ores range in ages from the Paleoproterozoic to the Mesoproterozoic. The Precambrian age of high-grade hematite deposits imposes even greater limitations on the use of the APWP method, because the Precambrian paleomagnetic database is relatively small. Relatively complete Phanerozoic APWP have led to better success, at present, in dating typically, much younger orebodies such as MVT deposits (e.g., Leach et al., 2001). Another factor to consider is the strong anisotropy of susceptibility observed in many BIF (e.g., Schmidt and Clark, 1994), which can confound paleomagnetic measurements and have led many workers to shy away from conducting studies on BIF-hosted iron ore deposits.

Despite these limitations, the current database of studies allows a view of high-grade hematite deposits within a global temporal framework. Hard high-grade hematite deposits show variable associations with major tectonic, erosional, and magmatic events, supporting the view that a single metallogenic model cannot explain these deposits. However, deposits do all share a Precambrian age and proposed models should account for this. It is acknowledged that the common age distribution of deposits should be tested, but this can be easily achieved, and at relatively low cost, by conducting paleomagnetic studies on those deposits for which no paleomagnetic data currently exist (i.e., the voluminous deposits of South America and India). There already exists a large database for these deposits that includes detailed geological settings and geochemical data. Furthermore, the continued updating of APWP for the Gawler, Pilbara, and Kaapvaal cratons may allow much better age resolution for those deposits that have been studied previously by the paleomagnetic method. Resampling of previously studied ore deposits, with the employment of modern paleomagnetic methods, will also add much-needed constraints to the current database.

If the predominantly or strictly Precambrian age distribution proves to be true, it would have very important implications toward exploration for new deposits. There is a distinct possibility that Phanerozoic deformational structures and unconformities as well as young magmatic fluid sources can largely be ignored in the search for hard ores. Caution is advised in the case of apparently younger structures such as faults, since these may be long-lived and multiply reactivated. They could have been active during the Precambrian and may have served as conduits for hydrothermal fluids responsible for ore formation. Potential multiple phases of ore formation or upgrading may be present, and these may be associated with magmatic or metamorphic fluid sources or be tectonically driven. Once potential hard ore targets have been identified, younger erosional surfaces or unconformities (of Precambrian or Phanerozoic age) that intersect them regain importance for potential development of friable or soft ore.

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