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Low paleointensities recorded in 1 to 2.4 Ga Proterozoic dykes, Superior Province, Canada☆

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Abstract

Paleointensity studies have been conducted on six mafic dyke swarms from the Superior Province in Canada with U–Pb ages between 1141 Ma and 2473 Ma (Buchan et al., Tectonophysics 319 (2000) 167–198). The mean direction of the characteristic magnetization for each dyke swarm coincides with results of earlier studies in which the primary origin of the magnetization was established on the basis of contact and secular variation tests. This primary component was isolated over a very narrow range of high unblocking temperatures (usually 550–580°C), indicating that it is carried by magnetite or low titanium titanomagnetite. Magnetic mineralogy and grain size experiments suggest that the magnetization is dominated by pseudo-single domain or single domain grains. Taken together, the characteristics of this component suggest that it is thermal in origin. Paleointensity experiments were conducted with a specially designed oven, using a revised approach of the Thellier–Coe method. Eighty-five successful determinations of paleointensity were obtained for 20 sites from six dyke swarms, doubling the number of existing well-constrained Precambrian data. The paleofield estimates for the dyke swarms vary between 5 and 11.2 μT, yielding virtual dipole moments between 0.85 ± 0.1 and $2.8 \pm 0.87 \times 10^{22}$ Am^2. The overall convergence between these new results and the rest of the database strongly reinforces the existence of a weaker geomagnetic field during the 1000–2400 Ma period than has been recorded between 0.3 and 400 Ma. However, additional determinations are required, particularly for the 400–2000 Ma interval, in order to establish whether a long-term evolution in the time-averaged field intensity can be linked to the onset of the growth of the Earth's inner core.

Keywords: paleointensity; Precambrian; rock magnetism; mafic dykes; Superior Province

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1. Introduction

The existence of the Earth's magnetic field as early as 3.5 Ga is supported by paleomagnetic data [1,2]. It also seems plausible that changes in the convection pattern of the core and related changes to the magnetic field could be caused by the growth of the solid inner core following either release of latent heat or chemical differentiation.

How and when the inner core grew remains uncertain. In a recent model for cooling of the Earth's core, Labrosse et al. [3] proposed that growth of the inner core cannot be older than 1.7 Ga. This age is in accordance with previous results [4,5] although it is assumed that excess heat is evacuated by a conductive shell at the top of the core rather than being dissipated by compositional convection. In a subsequent paper, Labrosse et al. [6] concluded that, in the absence of radioactive elements in the core, the formation of the inner core could not have started before 2.5 Ga and that a more realistic age for the initiation of the inner core is 1 ± 0.5 Ga. They further calculated that this age could be extended to 3 Ga if radioactive elements are present in the core.

Paleomagnetic experiments can help establish the timing of the growth of the inner core. According to the model developed by Hollerbach and Jones [7], before the formation of the inner core, the field would have been characterized by more excursions and greater paleosecular variation. However, recognizing such effects would require the acquisition of a large amount of paleomagnetic data. Following an early suggestion by Stevenson [4], a second possibility is that the formation of the inner core would cause a significant evolution in geomagnetic intensity. A compilation of paleointensity determinations from Archean and Proterozoic samples from the Kaapvaal and Superior cratons by Hale [8] was interpreted as being consistent with a sharp increase in geomagnetic strength between 2.7 and 2.1 Ga. However, this analysis relied on few data, most of which do not satisfy current quality criteria. Several recent studies [9–12] reported virtual dipole moments (VDM) for the Precambrian which are lower than the present-day field. However, there are also exceptions with relatively high VDM values [13]. In any case, the present Precambrian paleointensity database consists of a rather small number of determinations, some of which are of poor quality (Table 1).

The aim of this study was to improve the present database by acquiring new paleointensity results from six Canadian diabase dyke swarms, each with a 'key' paleopole as defined by Buchan et al. [14]. The choice of these formations was constrained by the existence of reliable radiogenic ages (from 1141 to 2473 Ma) and the primary origin of their magnetization.

2. Geological setting and sampling

The Superior Province of the Canadian Shield consists of an Archean assemblage of volcanic–plutonic and metasedimentary–gneissic subprovinces. It is the largest of the Archean blocks

Table 1
Previous Precambrian Thellier-type paleointensity determinations (since 1970)

Country	Rock formation	Age	*N*	*n*	M	Test	*F*	±	VDM	Ref.
Canada	Keeweenawan	1100 ± 100	21	54	T	C+	44	14	11.4 ± 3.6	[39]
Africa	Komati	3470 ± 20	–	4	T	M	25	10	4.69 ± 0.13	[57]
Canada	Tudor gabbro	1150 ± 50	9	19/45sp	T+	M	22.6	3.8	4.6 ± 0.8	[9]
Canada	8a dyke	ca. 2600	1	5/9sp	T+	C+	30.9	1.1	6.3 ± 0.2	[13]
Canada				2/6p	T+	–	43.9	1.2	9 ± 0.2	[13]
Australia	Hamersley basin	ca. 2000	3	8/22sp	T	F	6.88–13.94	2.4	2.4 ± 1	[12]
Greenland	Dolerite dyke	2752 ± 63	1	7/12sp	T+	C+	13.5	4.4	1.9 ± 0.6	[11]
USA	Stillwater Complex	2700	1	3	T+	M	31.7	5.75	4.1 ± 0.5	[10]

Column headings indicate: country, rock formation names, age, *N*: number of sites, *n*: number of samples, M: method (T: Thellier-type method, T+: Thellier-type method with pTRM checks), magnetic tests (C+: positive contact test, F: fold test), F: paleointensity in μT, ±: standard deviation in μT, VDM: virtual dipole moment in 10^{22} Am^2 and standard deviation, references.

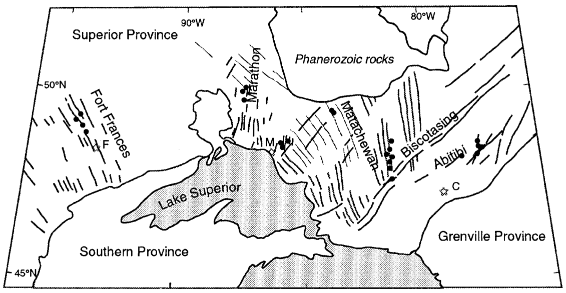


Fig. 1. Simplified geological map of the Superior Province (Canada) and location of dyke swarms and sites (modified from Buchan et al. [22]). M, C and F represent the towns of Marathon, Cobalt and Fort Frances, respectively.

that amalgamated during the Paleoproterozoic to form Laurentia [15]. Many giant dyke swarms with a variety of trends, average dyke thicknesses and compositions intruded the Superior Province during the Proterozoic. The dykes of some swarms, such as the Matachewan and Mistassini, exhibit a radiating distribution. They are referred to as giant radiating dyke swarms and have been linked to mantle plumes (e.g. [16]). The location of their foci along cratonic margins suggests an association with rifting and continental breakup.

A number of the dyke swarms of the Superior Province have been studied in detail in recent years, and now have good ages (e.g. [17–20]), as well as reliable paleomagnetism (e.g. [18,19,21–23]) and geochemistry analyses (e.g. [24,25]). In the course of these studies earlier interpretations of the age and distribution of swarms have been clarified. For example, northeast to north-northeast trending dykes of western Quebec were originally all referred to as Abitibi dykes (e.g. [26]), then later separated into two sets, the Abitibi and Preissac swarms (e.g. [27]). However, recent work has demonstrated that they belong to three swarms of distinctly different ages, the Abitibi swarm at 1141 Ma, the Biscotasing swarm at 2167 Ma and the Senneterre swarm at 2216 Ma (see summary in [18]).

We selected six dyke swarms (Fig. 1) for paleointensity study based on their stable paleomagnetic remanences and the existence of at least one U–Pb baddeleyite or zircon age determination (Table 2). The oldest swarm that was examined is

Table 2
Sampling sites

Dyke swarm	Age (Ma)	Ref. age	Plat	Plong	A95	Ref.
Matachewan	2473±16/9 ; 2446±3	[20]	-42	238	3	[21], [40]
SITE	N	W	trend	dip	T	Pal.
K1	48°33'23.3	81°37'0.49"	-	-	>5m	7°
K2	near K1		-	-	>5m	7°
GT1	48°09.32.4	81°35 14.7	N10	90	>20m	7°
GT2	near GT1		N358	90	>10m	7°
GT3	47°44'58.8	81°38'00.9	N350	90	>10	6°
GT4	47°44'49.2	81°39'33.4	N355	82E	12m	7°
GT5	47°41'58.4	81°46'26.5	N320	-	>100m	7°
GT6	47°39'57.6	81°47'31.3	N350	-	>18m	6°
H1	49°30'41.4	83°51'45.1	N	-	>3m	6°
H2	49°32'11.6	83°49'33.5	N	-	>20m	6°
Senneterre	2216+8/-4	[18], [19]	-15	284	6	[18]
Q5	48°22'54.3	78°00'13.4	NE	-	>5m	27°
Biscotasing	2167±2	[18]	28	223	8	[18]
GT7	-	-	-	-	>5m	41°
Q1	48°16'21.8	78°12'28.4	-	-	>16m	41°
Q10	near Q9		-	-	>100m	41°
Marathon	2121+14/-7	[19]	43	196	8	[19]
WL1	48°42'36.6	85°36'03.4	N30	75SE	15m	37°
WL2			N357	-	10m	37°
WL3	48°42'53.5	85°35'00.8	N10	89E	8m	37°
MA1	49°01425.2	85°53'42.5	N28	37E	>50m	37°
LL1	49°46'03.4	86°25'30.0	N0	-	35m	38°
LL2	near LL1		-	-	3m	38°
LL3	49°45'20.8	86°39'06.2	N10	-	>5m	39°
Fort Frances	2076+5/-4	[19]	43	184	6	[28]
KF1	49°08'27.0	93°56.42.2	-	-	>3m	35°
KF2	near KF1		N325	-	>10m	35°
KF3	nearKF4					36°
KF4	49°33'35.2	94°03'47.5	NNW	-	>15m	36°
Abitibi	1141±1	[17]	43	208	14	[23]
Q2	48°22'50.9	77°58'32.7	NE	-	>16m	39°
Q3	48°40'36.7	78°07'44.2	-	-	>3m	40°
Q4	48°17'43.7	78°11'01.1	-	-	>3m	40°
Q5	48°22'54.3	78°00'13.4	NE	-	>5m	40°
Q6	48°51'52.3	79°23'20.3	-	-	>5m	41°
Q7	48°53'19.7	79°30'17.0	NE	-	>10m	41°
Q8	100m Q7		-	-	>20m	41°
Q9	48°53'35.9	79°23'14.0	-	-	>3m	41°
T1	48°10'56.1	81°34'58.7	-	-		42°
T2	48°10'56.2	81°34'58.8	-	-	250 m	42°
T3	48°11'59.7	80°34'00.1	-	-		42°

Portions of the table with gray shading give information on the dyke swarms, with column headings indicating the following: dyke swarm name, age in Ma obtained by U–Pb method, age references, Plat, Plong and A_{95} are latitude, longitude and radius of the circle of 95% confidence of the paleomagnetic pole, and paleomagnetic references. Unshaded portions give information on sampling sites: site number, geographic coordinates (north and west), dyke trend, dip and thickness (T), and paleolatitude (Pal.) of the site.

the 2473–2446 Ma Matachewan swarm [20,21]. Four swarms have ages that fall in the 2216–2076 Ma period: namely the 2216+8/−4 Ma Senneterre swarm [18,19], the 2167 ± 4 Ma Biscotasing [18], the 2121+14/−7 Ma normally magnetized portion of the Marathon swarm [19] and the 2076 Ma Fort Frances swarm [19,28]. In addition, a single Mesoproterozoic swarm, the 1141 ± 1 Ma Abitibi swarm [17,23], was selected.

Previous paleomagnetic studies indicated that each swarm carries a stable magnetization with a narrow range of blocking temperatures near the Curie temperature of magnetite. Furthermore, positive baked contact tests are available to demonstrate that the Matachewan, Biscotasing and Abitibi remanences are primary [18,23,29].

The characteristic Senneterre and Fort Frances magnetizations are also known to date from the time of dyke swarm emplacement on the basis of 'secular variation' tests in which secular variation is detected between dykes that occur in the same geographic locality [18,28]. No field test has yet been carried out for the normally magnetized Marathon dykes. However, a positive baked contact test is available for the reversely magnetized Marathon dykes [19] which are only about 20 Myr younger than the Marathon normal dykes [30] and have a remanence direction that is only slightly different. This suggests that the Marathon normal remanence direction is also primary.

In general, Proterozoic dyke swarms of the Superior Province are largely undeformed and unmetamorphosed. Dykes are usually subvertical,

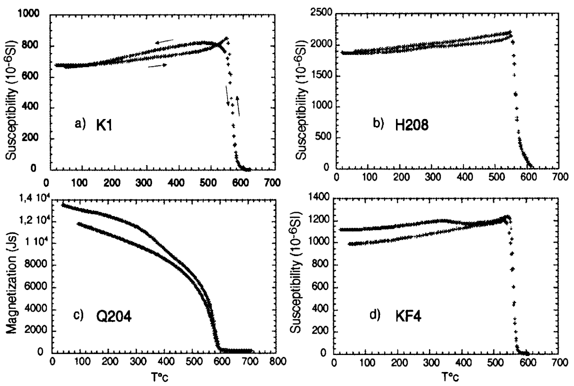


Fig. 2. Examples of thermomagnetic curves obtained in weak fields (a,b,d) and in high fields (c).

except in close proximity to the ca. 1100 Ma Mid-Continent Rift [29]. Likewise, the primary linear or radiating geometry of the swarms has been preserved, except in proximity to the Kapuskasing Structural Zone, a prominent tectonic feature which has distorted the radiating pattern of the Matachewan dyke swarm [21].

Some samples used in the present paleointensity experiments were obtained from the existing collections of the Geological Survey of Canada in Ottawa. In addition, we collected 300 cores from five or more dykes per swarm, with at least eight cores per site. Sampling was carried out using a portable gasoline-powered drill. The cores were oriented with a magnetic compass, and whenever possible, with a sun compass. Differences between sun and magnetic readings did not exceed 15°.

3. Magnetic mineralogy and granulometry

Rock magnetic experiments were conducted at the Parc St Maur IPGP laboratory in order to evaluate the suitability of the samples for paleointensity determinations. Weak-field thermomagnetic experiments (K–T) were performed on 25 specimens (at least four samples per dyke swarm) with a KLY-2 kappabridge. Four experiments were also processed in high field (M_s–T) with a Curie balance (Fig. 2c). More than 80% of the experiments yielded reversible cooling and heating curves, as shown in Fig. 2a,b,d and Fig. 2c for low and high field experiments, respectively. Both M_s–T and K–T curves reveal a Curie point between 565°C and 580°C, consistent with the presence of either pure magnetite or low titanium titanomagnetite. The shapes of the heating and cooling curves are also characteristic of magnetite [31,32]. In some cases, a small bump appears in the cooling curve beyond 300°C (Fig. 2c), indicating the presence (or creation) of a middle-blocking temperature mineral phase, which disappears during subsequent heating. A striking feature of the K–T curves is the monotonic increase of low field susceptibility (with the exception of the small bump described above) up to a temperature as high as 540°C. This trend is followed by a Hopkinson peak which precedes a very rapid decrease

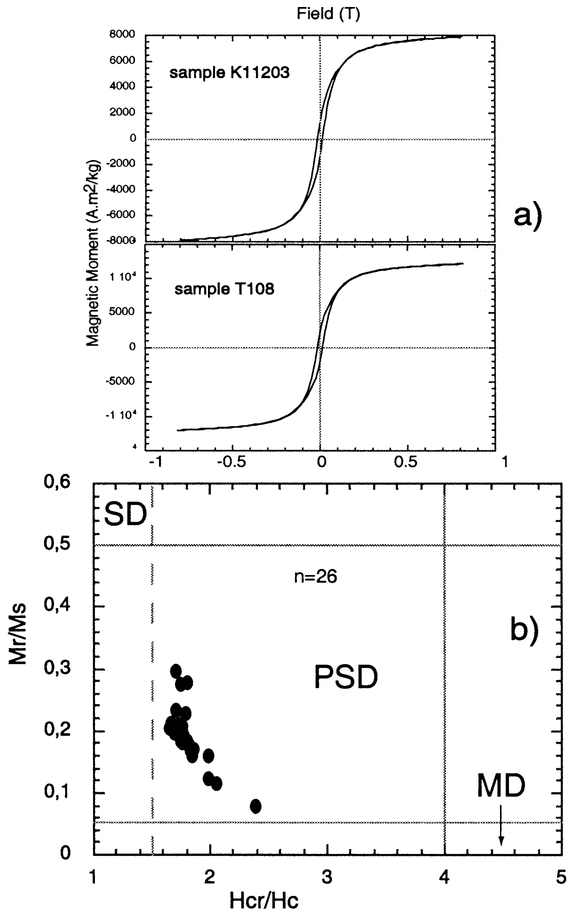


Fig. 3. (a) Typical examples of hysteresis curves (uncorrected) of small chip samples. (b) Hysteresis parameters from high field cycles (diagram after Day et al. [34]). Ratio of coercivity of remanence (Hcr) to coercivity (Hc) is plotted against the ratio of remanent magnetization (Mr) to saturation magnetization (Ms). SD: single domain; PSD: pseudo-single domain; MD: multidomain; n, number of data.

of the susceptibility to the Curie temperature. The same critical temperature is associated with the M_s–T curves but in this case about two thirds of saturation magnetization has been lost previously. These curves are almost identical to those characteristic of pure synthetic magnetite [33] and thus probably indicate the presence of small grains of magnetite.

Hysteresis cycles were conducted on 28 samples using the vibrating magnetometer of the Parc St Maur laboratory. Typical examples are shown in Fig. 3a,b. They are all similar without any waspwaisted or potbellied loops. Hysteresis parameters calculated after correcting for para- and diamagnetism are shown on a Day diagram [34] in Fig. 3b. They are consistent with pseudo-single domain grain sizes but the presence of a mixture of monodomain and multidomain grains cannot be excluded. Hodych [35] reported the presence of intergrowth of nearly pure pseudo-single domain magnetite with ilmenite lamellae from scanning electron microscopic observations on samples from the Matachewan and the Biscotasing dyke swarms. These samples were acquired from the same dyke collection at the Geological Survey of Canada from which some of our samples were obtained. All these results converge to indicate that the major magnetic phase in the dyke swarms under study consists of small grains of magnetite.

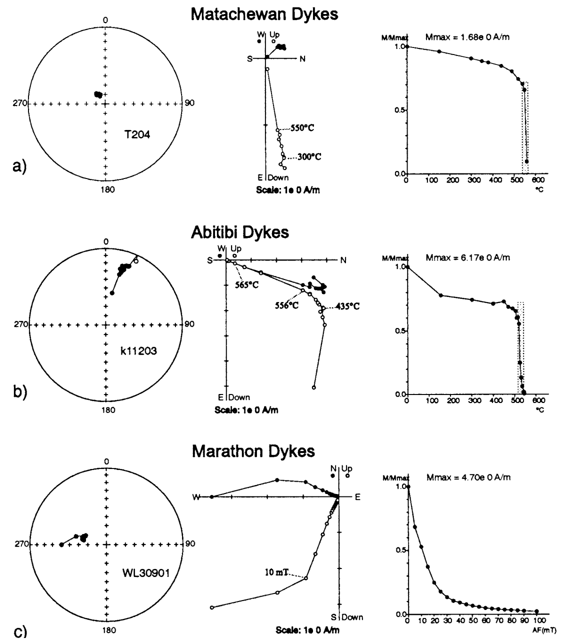


Fig. 4. Examples of thermal demagnetization (a and b) and alternating field (AF) demagnetization (c). For each sample the following are shown: a stereographic projection, an orthogonal vector plot with solid (open) symbols for data projected onto the horizontal (vertical) plane, and a normalized plot of the intensity of magnetization versus temperature or peak of AF.

4. Paleomagnetic components

Seventy-five samples (one to three samples per site) were thermally demagnetized in zero field (Fig. 4a,b). In addition, 38 samples were subjected to alternating fields (AF) (Fig. 4c) while 12 others were saturation magnetization stepwise demagnetized to 20 mT before being subjected to thermal cleaning. All measurements and demagnetization experiments were performed in the IPGP shielded room using a Pyrox oven, a 2G demagnetizer and a JR-5 spinner magnetometer. The objective of this preliminary study was to identify sites that were unsuitable for paleointensity experiments and to constrain the unblocking temperature spectrum of the characteristic remanent magnetization (ChRM).

Typical demagnetization diagrams obtained with both thermal and AF techniques are shown in Fig. 4. All samples display a succession of two or three components. The first one with an erratic direction was easily removed by heating at 100°C and thus was probably acquired during storage of the samples. The second component could be very resistant to thermal demagnetization and persisted up to temperatures of at least 300°C, and in some instances to 540–550°C. In contrast it was generally much less resistant to alternating fields and could be removed by peak fields as low as 20 mT. Given the homogeneous character of the magnetic mineralogy as indicated by the narrow range of Curie temperatures, this component is most likely not related to a distinct mineralogical phase but rather to granulometry. Little resistance to AF demagnetization combined with strong resistance to thermal demagnetization is typical behavior of multidomain grains. The rock magnetic characteristics described above are consistent with this interpretation, but the exact origin of this medium high temperature component remains difficult to determine because of overlapping unblocking temperatures with those of the third component, the ChRM. In some cases, the medium high temperature component has a direction similar to that of the present Earth's magnetic field, suggesting recent acquisition due to viscous or thermoviscous processes [19,37]. In others, the direction of this component may indicate acquisition as an overprint during the Proterozoic [36,38].

In almost all cases, the ChRM was isolated above 550°C (Fig. 4). Therefore, the range of ChRM unblocking temperatures did not exceed 30°C, and in some specimens 50% of the magnetization was lost over an interval as narrow as 5°C. Note that similar characteristics have been reported for the Modipe Gabbro in Africa [2] and the Stillwater Complex [10]. The high temperature pattern of the demagnetization diagrams is very similar to that of the thermomagnetic curves described above and confirms that the ChRM is carried by pure magnetite or low titanium titanomagnetite. Such a high unblocking temperature spectrum is associated with relaxation times as long as tens of billions of years [39]. Thus these samples carry a very stable magnetization over geological time.

As noted earlier, the primary origin of the high temperature (ChRM) component was established, in most cases, on the basis of positive contact tests. This means that magnetization was acquired at the time of dyke emplacement by cooling from the intrusive magma temperature to that of the host rocks. The paleomagnetic poles for each dyke swarm were summarized by Buchan et al. [14]. They are tabulated in Table 2 along with the original references [18,19,40,41]. All are considered to be 'key' paleopoles, a term that identified poles that are well-defined and precisely dated. They represent the most robust paleomagnetic data that have been obtained so far for this period. Using this data set of paleopoles as a reference, the measurements of the pilot specimens allowed us to identify sites with inconsistent directions. All sites, except Q9, gave suitable directions.

The directions of high temperature (ChRM) components are shown in Fig. 5. For each dyke swarm, we plot the site mean directions, the swarm mean direction and the expected direction, which was calculated at a mean locality using the 'key pole' (e.g. [14]) (Table 2). Except for the Senneterre dyke (with only one site represented), the swarm mean directions do not differ more, from the expected directions, than expected for secular variation. Short-term variations of the

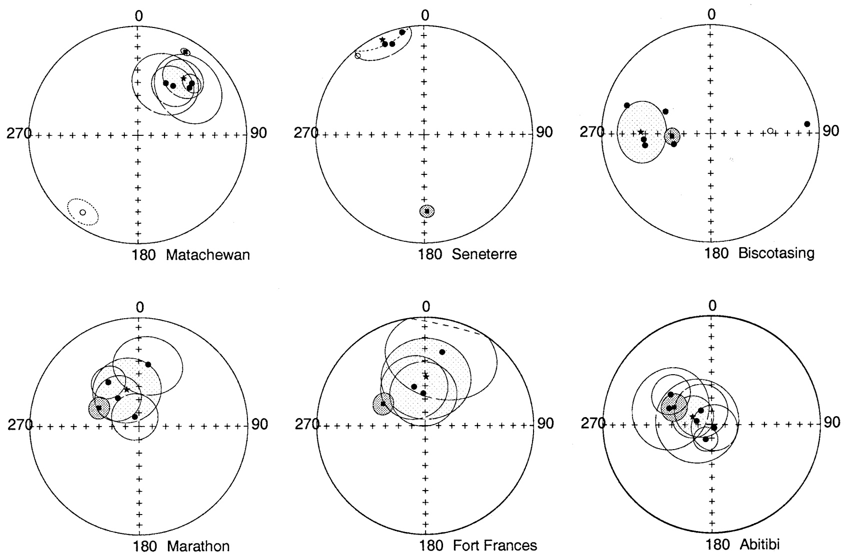


Fig. 5. Stereographic projections of mean paleomagnetic directions. For the Matachewan, Marathon, Fort Frances and Abitibi dyke swarms, circles represent site means and the star represents the dyke swarm mean. Squares are the projection of the expected paleomagnetic directions for each swarm. For the Senneterre and the Biscotasing dyke swarms, in which only one site is represented, circles are individual samples and the star is the mean direction for the site. Solid (open) symbols represent positive (negative) inclination.

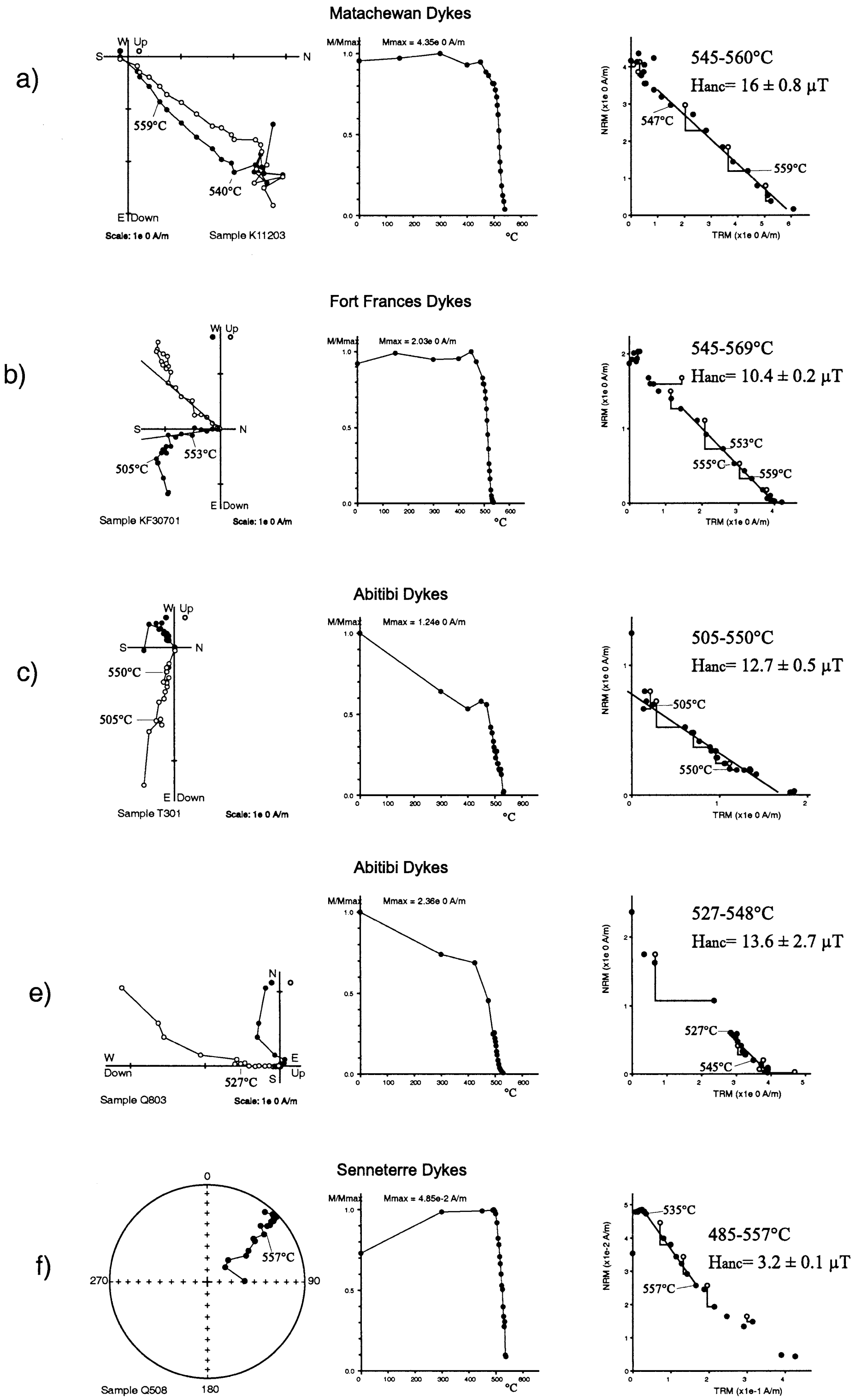
magnetic field seem to be averaged for the 2.4 Ga Matachewan and 1.1 Ga Abitibi dyke swarm, but not for the four ca. 2.2–2.0 Ga swarms for which there were fewer samples per site as well as fewer dykes per swarm.

In summary, the high temperature ChRM component of the dyke swarms is suitable for paleointensity study. Positive contact or secular variation tests for each of the six swarms have been reported in the literature, demonstrating that this highly stable remanence is primary in origin. Rock magnetic experiments herein on samples from the six swarms and electron microscope observations by Hodych [35] on samples from two of the swarms suggest that the remanence carrier in these rocks is single domain or pseudo-single domain magnetite or low titanium titanomagnetite. Such fine-grained magnetite is thought to be a likely carrier of thermoremanent magnetization (TRM) in igneous rocks. Thus, the ChRM component is undoubtedly primary and likely a TRM acquired during initial cooling of the dykes.

5. Absolute paleointensity

5.1. Experimental protocol and technical aspects

Paleointensity experiments were conducted using the Coe version of the Thellier and Thellier [42] experiments but with a different protocol. Coe [43] proposed demagnetizing the natural remanent magnetization (NRM) in zero field and then producing a laboratory TRM. However, measuring the NRM first does not allow one to detect remagnetization components that can be produced in zero field with unblocking temperatures higher than the last temperature step T_i. During acquisition of the subsequent TRM(T_i), the same grains will keep a zero magnetization. They will evidently be involved in the magnetization acquired during the following step (T_{i+1}), but remain undetectable and ultimately yield incorrect paleofield determinations. In order to solve this problem, we imparted the TRM before heating the sample in zero field [44,45]. In this case magnetomineralogical transformations occur in the


Matachewan Dykes
a)
W Up
S
N
559°C
540°C
E Down
Scale: 1e 0 A/m
Sample K11203
M/Mmax
Mmax = 4.35e 0 A/m
°C
545-560°C
Hanc= 16 ± 0.8 μT
547°C
559°C
NRM (x1e 0 A/m)
TRM (x1e 0 A/m)
Fort Frances Dykes
b)
W Up
S
N
553°C
505°C
E Down
Sample KF30701
Scale: 1e 0 A/m
M/Mmax
Mmax = 2.03e 0 A/m
°C
545-569°C
Hanc= 10.4 ± 0.2 μT
553°C
555°C
559°C
NRM (x1e 0 A/m)
TRM (x1e 0 A/m)
Abitibi Dykes
c)
W Up
S
N
550°C
505°C
E Down
Sample T301
Scale: 1e 0 A/m
M/Mmax
Mmax = 1.24e 0 A/m
°C
505-550°C
Hanc= 12.7 ± 0.5 μT
505°C
550°C
NRM (x1e 0 A/m)
TRM (x1e 0 A/m)
Abitibi Dykes
e)
N
W
Down
E
Up
527°C
S
Sample Q803
Scale: 1e 0 A/m
M/Mmax
Mmax = 2.36e 0 A/m
°C
527-548°C
Hanc= 13.6 ± 2.7 μT
527°C
545°C
NRM (x1e 0 A/m)
TRM (x1e 0 A/m)
Senneterre Dykes
f)
0
270
90
180
557°C
Sample Q508
M/Mmax
Mmax = 4.85e-2 A/m
°C
535°C
485-557°C
Hanc= 3.2 ± 0.1 μT
557°C
NRM (x1e-2 A/m)
TRM (x1e-1 A/m)

presence of the field and thus result in a chemical remanent magnetization (CRM) component that is immediately detected by a deviation of the NRM toward the direction of the field in the oven. During this procedure, partial TRM (pTRM) checks were regularly performed, mostly at T_{i-3}.

We conducted two series of experiments. The first one involved 15 double heating steps on standard cylindrical samples in a Pyrox paleointensity furnace with at least eight pTRM checks. Due to the very narrow range (about 30°C) of the unblocking temperatures of the ChRM, high temperature increments had to be 5°C or less to constrain the decrease of the NRM. In other words, reproducibility of repeated heating steps at the same temperature had to be within 2°C. Due to its large volume, the paleointensity oven used in the first series of experiments did not provide sufficiently accurate temperature control. Therefore, a new furnace was designed and built with an inner diameter of only 4 cm. The ends were closed with plugs in order to extend the length of the effective heating zone to 20 cm. Samples were separated in the furnace by a minimum of 0.5 cm. Two thermocouples located within the heating elements close to the samples allowed us to operate at temperature intervals as small as 1.5°C. We used 1 cm^3 specimens in order to optimize the number that could be placed in the furnace for each experimental run. A total of 35–40 samples were processed during each experiment and measured with a 2G horizontal cryogenic magnetometer in the shielded room of the IPGP paleomagnetic laboratory. These new experiments involved 25 double heatings and 10 pTRM checks. Typically, we performed six steps between 150 and 450°C (150, 300, 400, 450, 475, 500°C) and then 5°C or 3.5°C steps to 550°C. The final increments between 550°C and 572°C were reduced to 1.5°C or 2°C. As a consequence, most samples underwent a total of 60 h of heating–cooling cycles.

5.2. NRM–TRM (Arai) plots

Demagnetization diagrams, graphs of magnetic moment as a function of temperature and Arai plots [46] (in which the NRM remaining at each temperature step is plotted against the TRM gained in the laboratory field) have been systematically used to scrutinize the results of the paleointensity experiments (Fig. 6). The three plots provide an overview of the relationship between changes in direction and moment of the vector. The evolution of the NRM directions in sample coordinates gives us crucial information about possible deviations caused by CRM with blocking temperatures higher than the last heating. This is reflected by a tendency of the residual NRM to move towards the direction of the applied field. The demagnetization diagrams are also very important in constraining the temperature range over which the ChRM has been isolated.

One can distinguish Arai plots with a single linear slope from those with more complex behavior. A characteristic common to the first category is that the slopes of the best fitting lines are essentially restricted to a very narrow range of unblocking temperatures. This was expected on the basis of the thermomagnetic and demagnetization characteristics. All data points are perfectly aligned along the fitting line obtained by least square regression. The only minor deviation that is observed is caused by a low temperature pTRM check which actually (Fig. 6a,b) coincides with the small bump observed in the thermomagnetic diagrams. Note that this deviation occurred before the primary component was isolated. There was no indication that the ChRM was affected or that the NRM was partly remagnetized. A second small deviation is observed in some samples at intermediate temperatures ($<$ 500°C). In the second category of Arai plots, only a segment of the total NRM decrease was involved in the determination of the slope. This was caused either by the presence of a low intermediate temperature component (Fig. 6c,d) or by the occurrence of irre-

←
Fig. 6. Examples of successful intensity determinations. For each sample an orthogonal vector plot (solid (open) symbols represent positive (negative) inclination), a stereographic projection and an Arai diagram are shown.

versible changes with negative pTRM checks at very high temperatures (Fig. 6e). Several additional experiments were performed in order to check whether changes in magnetization would have remained undetected, and to investigate further the primary origin of the ChRM.

5.3. Additional checks

Fifteen samples previously used for the paleointensity experiments were remagnetized by heating at 600°C in a 22 μT field before repeating a complete new paleointensity protocol. Except for sample GT30101, all results provided the expected field value of 22 μT (Fig. 7a,b). If mineralogical changes occurred during initial Thellier experiments, they should induce incorrect determinations of paleointensity. This hypothesis would exclude that the same characteristics be repeated during the two experiments.

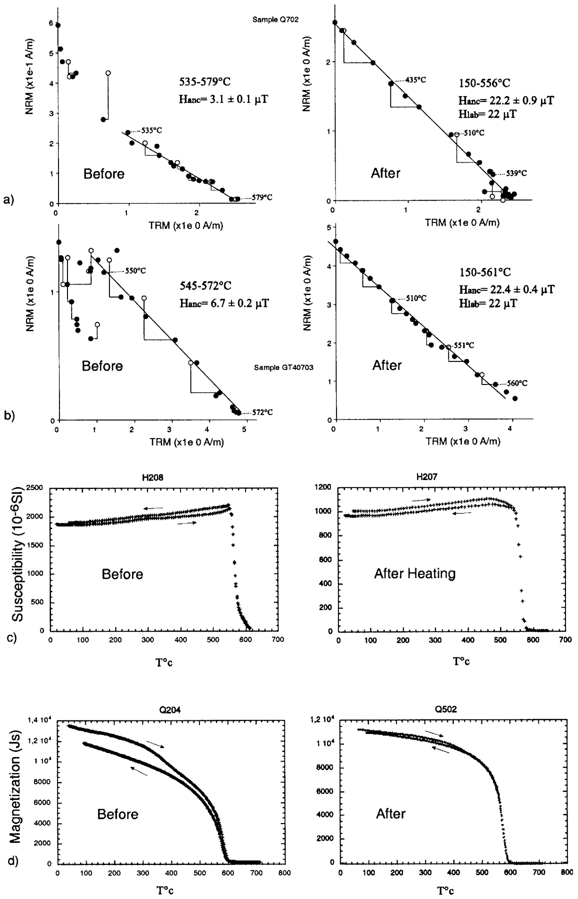


Fig. 7. Comparison of magnetic properties before and after paleointensity experiments (samples reheated at 600°C and 22 μT). Arai diagrams (a,b). Thermomagnetic evolution with temperature in weak (c) and in strong fields (d).

The rock magnetic properties can be compared before and after the paleointensity experiments using the thermomagnetic curves. The results obtained in weak (Fig. 7c) and strong (Fig. 7d) fields do not reveal any difference in magnetic mineralogy with respect to the pre-heating curves. Following a similar approach, the repeated hysteresis parameters of 21 measurements (Fig. 8) confirm that the properties remained basically the same before and after heating. There is no significant evolution in magnetic grain sizes, which remain dominated by pseudo-monodomains. The hysteresis loops also remain unchanged. From all these positive checks, we infer that the isolated negative checks were most likely caused by inaccuracies in temperature control.

The 250 m wide Abitibi dyke located near the city of Timmins provided the opportunity of testing effect of cooling rate on the acquisition of magnetization. Sampling sites were located in the two fine-grained dyke margins (petrological observations) (sites T1 and T3) and in the coarse-grained (T2) central part of the dyke to compare the influence of differential cooling. Sites T2 and T3 yielded very similar paleointensity estimates of 12.3 ± 2.3 μT and 10.7 ± 3.6 μT, respectively. Site T1 gave a lower value (5.3 ± 1.4 μT) but only three determinations could be obtained. In any case the results of T3 and T2 can be considered identical, which excludes any influence of the cooling rate.

5.4. Determination of absolute paleointensities

Paleointensities were calculated for samples whose Arai diagrams displayed a linear segment with a minimum of four data points in the temperature range of the ChRM and a correlation coefficient higher than 0.98. The fraction of the total NRM (factor *f* [47]) involved in this segment was at least 20%, and usually more than 50%. The pTRM checks were considered to be positive when they did not deviate by more than 10%

from the original TRM. Only one negative pTRM check over the ChRM temperature range was accepted. This small deviation with respect to the usual selection criteria (100% positive pTRM checks are usually required) was accepted because of the very small temperature intervals involved in the determination of the slope. Typical examples of unsuccessful experiments are shown in Fig. 9a,b. All samples characterized by concave-up or concave-down TRM–NRM curves were rejected. In addition, sites GT2, GT5, and GT6 were rejected because the primary component could not be correctly separated.

Due to uncertainties generated by insufficient temperature control in the original oven we decided to reject the entire data set obtained in the first series of experiments. Using the new furnace, we obtained 85 successful determinations from a total of 220 experiments. The results, listed in

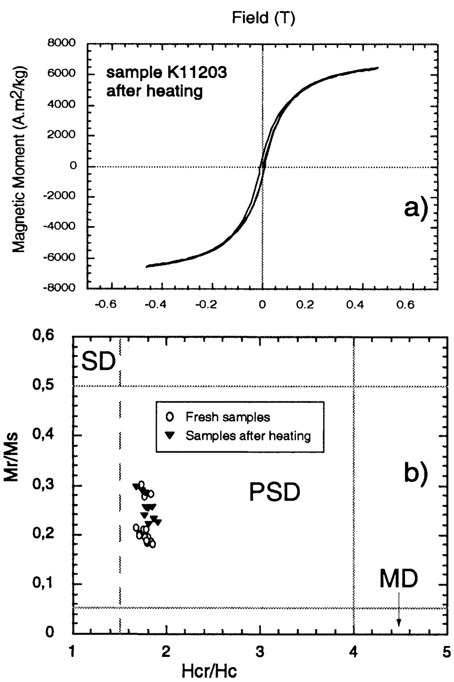


Fig. 8. (a) Typical example of hysteresis curves (uncorrected) of a small chip sample after paleointensity experiment. (b) Comparison of hysteresis parameters for fresh (open circles) and heated samples (dark triangles) (diagram after Day et al. [34]).

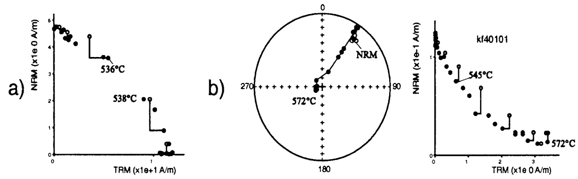


Fig. 9. Examples of rejected Arai diagrams.

Table 3[1], indicate the ancient paleofield, the standard error and the *f*, *g*, *q* [47] and *w* [48] quality factors. The paleofield intensities fall between 2 and 18.3 μT for individual samples and between 3.4 and 15.4 μT for the site means. We also calculated the average paleointensity of each dyke swarm (Table 3[1]).

6. Discussion

6.1. Dominant field geometry during the Precambrian

The calculation of a dipole moment relies on the assumption that the Precambrian field was dominantly dipolar. Several studies have attempted to document the field geometry during this remote period. In an early study, Evans [49] took advantage of the fact that each zonal multipole component exhibits a specific relation between inclination and latitude. Thus, the frequency distribution of the paleomagnetic inclinations can be tested against the theoretical distribution of each zonal field component. An underlying hypothesis is that plate motions have been large enough over periods as long as 10^8–10^9 years to randomize the spatial sampling of the field. Evans [49] did not detect any significant departure from the distribution predicted by an axial dipole during the Phanerozoic. More than 10 years later, Piper and Grant [50] confirmed this result, but suggested some deviation from the geomagnetic axial dipole (GAD) for older periods. Recently, Kent and Smethurst [51] applied the same analysis to the global paleomagnetic data-

[1] See Table 3 in the online version of this paper.

base [52]. They observed that the Cenozoic (0–65 Ma) and Mesozoic (65–250 Ma) data are still consistent with the GAD but that the Paleozoic (250–550 Ma) and Precambrian (550–3500 Ma) results exhibit lower magnetic inclinations than expected from an axial dipolar distribution. Kent and Smethurst do not exclude the possibility that low magnetic inclinations during the Precambrian may result from the cratons having occupied low latitudes. Nevertheless, they proposed that 25% zonal octupole and 10% zonal quadrupole contributions be added to the axial dipole. This yields underestimates of the mid-paleolatitudes by up to 15° when the dipole-only hypothesis is used. Under this assumption, the VDMs would be reduced by up to 25%. The octupole–quadrupole contribution derived from the model proposed by Kent and Smethurst also implies the existence of asymmetrical polarities. Pesonen and Halls [36] reported two or three asymmetric reversals from ca. 1100 Ma Keeweenawan rocks. However, late Mesoproterozoic sediments from the Siberian craton yielded at least 15 symmetric geomagnetic reversals [53]. In addition, Gallet et al. [53] reported consistent paleomagnetic directions in two coeval sedimentary formations that were deposited between 1050 and 1100 Ma and separated by several thousand kilometers. They report positive reversal, fold and conglomerate tests, which suggest that the remanences are primary. Therefore, these results suggest a persistent dominant dipolar character of the field at that time. In this paper, we base our calculations on the GAD model for the Earth's magnetic field. However, it is clear that much more study of the Precambrian magnetic field is required to adequately test this model.

6.2. Precambrian field intensity

The paleolatitudes of the sampling sites were derived from the mean inclinations using the mean key poles [14] (Table 2). The mean VDM of each dyke swarm was calculated with reference to the paleopositions. The values are summarized in Table 4.

The results were assigned to the 'A' category when defined by at least three determinations for each site. The Matachewan and the Abitibi dykes belong to this category. The low Matachewan VDM value (2.8×10^{22} Am^2) was obtained from five sites with a total of 26 individual determinations. Similarly, the Abitibi VDM ($1.35\pm0.44\times10^{22}$ Am^2) is rather low. The other paleointensity results from the Senneterre, Biscotasing, Marathon and Fort Frances swarms are also considered reliable, but a little less accurate than the 'A' data because of the lower number of determinations at each site. They were assigned to the 'B' category and have yielded mean VDM values between $0.85\pm0.1\times10^{22}$ and $1.26\pm0.67\ \times10^{22}$ Am^2. In Fig. 10a, we have plotted all mean VDMs, for each dyke or site as a function of age. Note that these new determinations double the number of reliable data that are available for the 3.5–1.0 Ga period.

In Fig. 10b, all site-averaged VDMs have been plotted as a function of their age and compared with those of the existing database. Schwarz and

Table 4
Dyke swarm mean paleointensity results

Swarm	Age (Ma)	Sites	Samples	H (μT)	Error (μT)	VDM (10^{22} Am^2)	Error (10^{22} Am^2)
A mean intensity							
Matachewan	$2473\pm16/9$; 2446 ± 3	5	26	11	3.4	2.8	0.87
Abitibi	1141 ± 1	5	23	7.9	2.6	1.35	0.44
B mean intensity							
Senneterre	$2216\pm8/4$	1	4	6.2	3.3	1.26	0.67
Biscotasing	2167 ± 2	2	6	4.5	1.14	0.85	0.1
Marathon	$2121\pm14/7$	4	12	5.9	2.1	1.03	0.22
Fort Frances	$2076\pm5/4$	3	7	5.66	3.28	1.07	0.16

Column headings indicate: swarm name, age in Ma, number of sites, number of samples, H (mean paleointensity in μT), error (standard deviation), VDM (in 10^{22} Am^2) and its error.

Symons [54], Bergh [55] and Kobayashi [56] were the first to study the Precambrian field intensity from Canadian rocks. McElhinny and Evans [2] investigated the Modipe gabbro and the Hart dolerite in Africa. However, we prefer to discard these early results because they were obtained without pTRM checks. Pesonen and Halls [36] studied Keeweenawan dykes, sills and volcanics in Canada and obtained a dipole intensity that was about 45% higher than the present-day field. However, this study was performed without pTRM checks and the average field is characterized by very large dispersion. It is included in our discussions and in Fig. 10a,b because of the unusually large number of paleointensity determinations that were made. The oldest (3.5 Ga) paleofield (4.69×10^{22} Am^2) was obtained by Hale and Dunlop [57] and Hale [1] from the Komati Formation.

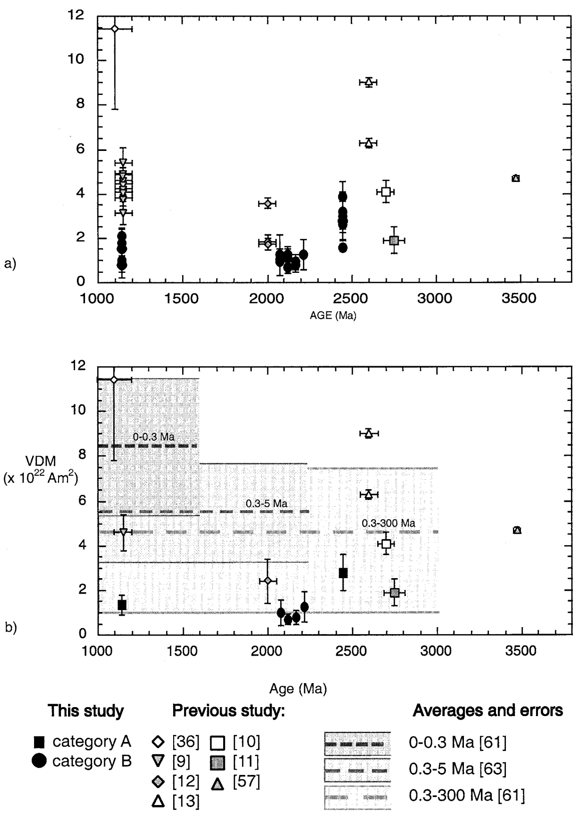


Fig. 10. Variation of VDM with time for the 1–3.5 Ga period.

During the last decade, several studies were performed using the Thellier–Coe technique with pTRM checks. They are shown in Fig. 10b. Except for the 8a dykes from the Slave Province [13], all of these recent results [9–12] are characterized by rather low mean VDMs which vary between 1.9 and 4.6×10^{22} Am^2.

6.3. Long-term evolution

The present results are next discussed within the framework of the existing paleointensity database for the whole geological record (PINT99) [58–60]. Since we are exploring very long-term field variations in geomagnetic intensity, we must deal with time-averaged values over sufficiently long periods of time. Selkin and Tauxe [61] combined the PINT99 database and the Scripps submarine basaltic glass database (SBG99) in order to extract a long-term evolution of the time-averaged field intensity. They found that the heavily sampled 0–0.3 Ma period has a mean VADM of $8.4\pm3.1\times10^{22}$ Am^2. Valet [62] selected only the Thellier determinations of the database and obtained a nearly identical mean VADM of $7.0\pm3.6\times10^{22}$ Am^2 for the past 4 million years and a mean value of $7.9\pm1.8\times10^{22}$ Am^2 from the sedimentary records calibrated with volcanic data. Using the Thellier determinations of the PINT99 database to which they added a large set of determinations from submarine basaltic glasses, Juarez and Tauxe [63] obtained an average dipole moment of $5.5\pm2.3\times10^{22}$ Am^2 for the 0.3–5 Ma interval. The mean intensity given by Selkin and Tauxe for the 0.3–300 Ma interval is $4.6\pm3.2\times10^{22}$ Am^2. Note that there is no significant difference between any of those means at the 1σ level, because of the large sizes of the standard deviations. On the other hand, the sampling density of 516 data points per Myr within the 0–0.3 Ma interval is 200-fold larger than during the 0.3–300 Ma interval (2.2 data per Myr). In other words, during the 0.3–300 Ma interval any period as long as the past 0.3 Myr would not be described on average by more than 0.75 data points. Thus, we must be very careful when comparing averaged field values for different time periods. Another major problem is that the choice of the boundaries that delineate the time-averaged intervals is very sensitive to the sampling distribution, which is highly irregular.

Although we are aware of these limitations, we can tentatively draw some conclusions from the overall record. In Fig. 10b, we show the averaged VDM values for the time periods described above. It is clear that most Precambrian VDMs lie below all the previous mean values, particularly below that for the 0.3–300 Ma interval. The VDM values obtained in the present study show relatively little dispersion. Most are in good agreement with earlier reliable results from the Precambrian, although there are also a few periods with large dispersion. The most noticeable time with larger scatter is around 1100 Ma. The Keeweenawan intrusions and lavas [36] and the Abitibi dyke swarm (this study) yield VDM values between 1.36 and 11×10^{22} Am^2. Given the very large error bars associated with the Keeweenawan data and the puzzling asymmetry in directions that were derived from that study, these data are still too scarce to drawn any valid conclusions.

All VDM values prior to ca. 1100 Ma indicate low field intensities, except for the two dikes (9 and 6×10^{22} Am^2) of the Yellowknife greenstone belt of the Slave Province studied by Yoshihara and Hamano [13]. We consider this study to be reliable as it meets all selection criteria (except for one negative contact test).

The number of paleointensity results in the 3.5–1.0 Ga period (Fig. 10b) is too small to allow any firm conclusions about the long-term evolution of the Earth's magnetic field. However, the dramatic rise in paleointensity between 2.7 and 2.1 Ga that was proposed by Hale [8] based on early and less reliable paleointensities is not supported by the data of Fig. 10b. In contrast, rather low paleointensities appear to dominate the 2.5–2.0 Ga interval. In particular, our B category estimates with ages between 2.07 and 2.21 Ga yield consistent low values. If we add the ca. 2.0 Ga value from Australia [12], they give a good picture of the 2.0–2.2 Ga field over a range of paleolatitudes. The average VDM for this period, $1.1\pm0.2\times10^{22}$ Am^2, is significantly lower than results obtained so far for the so-called Mesozoic dipole low (ca. 4×10^{22} Am^2) [60,64].

Additional paleointensity results for the 1.0–2.0 Ga period are required to improve the temporal distribution of the data. However, there is a possibly even more urgent need for data from the 0.4–1.0 Ga interval.

7. Conclusion

Several important conclusions can be drawn from this study:

1. Precambrian mafic dyke swarms from Canada are ideally suited for paleointensity study because they often carry a very stable magnetization which has been demonstrated to be primary by rigorous field tests. In addition, the magnetic mineralogy associated with this primary magnetization is relatively simple, consisting largely of single or pseudo-single domain grains of magnetite or low titanium titanomagnetite, suggesting that the primary remanence is most likely a TRM.
2. Eighty-five successful paleointensity determinations were obtained from six dyke swarms ranging in age from 2473 to 1114 Ma using a furnace that was designed specially to demagnetize samples with a narrow range of blocking temperatures and to allow temperature steps as small as 1.5°C. This new data set increases the existing paleointensity database for this period by a factor of two.
3. The VDMs from the six dyke swarms range between 0.85 ± 0.1 and 2.8×10^{22} Am^2. All are below the mean value of 4.6×10^{22} Am^2 obtained from other studies for the 0.3–300 Ma interval.
4. The data in this study, combined with other reliable data from the 3.5–1.0 Ga period, do not support the model proposed by Hale [8] of a dramatic rise in paleointensity in the 2.7–2.1 Ga period. Rather, a relatively low paleointensity appears to dominate at least in the latter portion of this interval (2.0–2.2 Ga). Only data in the early part of the period, obtained from dykes in the Slave Province at ca. 2.6 Ga, show a high paleointensity.
5. There are currently few reliable paleointensity data over much of the 0.4–2.0 Ga period. More paleointensity results of this age are

needed before a clear picture of the evolution of the Earth's magnetic field can be obtained, or its possible relationship to growth of the Earth's solid inner core can be discussed. The first order impression given by the overall still very scarce data set is that of an increasing dipole moment between 3 Ga and the present, but with such large standard deviations that this increase cannot yet safely be extracted from the data.

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