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Absolute paleointensity at 1.27 Ga from the Mackenzie dyke swarm (Canada)

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[1] Paleointensity studies have been conducted on 6 mafic dykes from the 1270 Ma Mackenzie swarm in the Slave Province (Canada). The mean direction of the characteristic magnetization coincides with results of an earlier study in which the primary origin of the magnetization was established on the basis of a contact test. High unblocking temperatures, magnetic mineralogy, and grain-size experiments suggest that the magnetization is dominated by pseudo-single domain or single domain grains of magnetite. Paleointensity experiments were conducted with a specially designed oven, using a revised version of the Thellier-Coe method. Thirteen successful determinations of paleointensity were obtained for 4 dykes. The paleofield estimates vary between 4.3 and 22.1 μ T, yielding virtual dipole moments (VDMs) between 1.3 ± 0.2 and $4.5 \pm 0.9 \times 10^{22}$ Am². These new results increase the number of low field determinations during the Precambrian, which largely dominate the database with an averaged field of $3.1 \pm 2.5 \times 10^{22}$ Am². They also emphasize the importance of additional studies to understand the differences with the strong paleointensities obtained using new techniques.

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Theme: Geomagnetic Field Behavior Over the Past 5 Myr

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1. Introduction

[2] According to the remanent magnetization carried by some rocks from Africa which provided the

oldest paleomagnetic records [Hale and Dunlop, 1984; McElhinny and Evans, 1968], the Earth's magnetic field was already active 3.5 billion years ago. During the past few years several studies were

conducted to improve the paleointensity database during the Precambrian periods. The present database was recently analyzed by *Dunlop and Yu* [2004], who reported that the published Virtual Dipole Moments (VDMs) lie within the range 0.5–1.5 times the 0.3–300 Ma average. A noticeable fact is that there is a majority of low field values, as indicated by the mean VDMs of $3.21 \pm 2.2 \times 10^{-22}$ Am² derived from the entire database or by the value of $3 \pm 1 \times 10^{-22}$ Am² if we restrain the calculation to the subset of determinations obtained by Thellier experiments including pTRM checks. Thus there is some indication that the field intensity was lower during the Precambrian than for the past 0.3–300 Ma. However, the distribution of the data is not Gaussian and there are also some high VDMs [*Smirnov et al.*, 2003; *Yoshihara and Hamano*, 2000; *Thomas and Piper*, 1995; *Thomas*, 1993] which remain to be explained. Such dispersion may be inherent to the overall field variability although some determinations from the same period can differ significantly. The aim of the present study was to add new results and check their consistency with the previous determinations. If we assume that the low mean field value is significant for this period, it is crucial to constrain its duration as it can have important implications regarding the evolution of the Earth's core and/or long-term changes at the core mantle boundary [*Labrosse and Macouin*, 2003; *Stevenson et al.*, 1983].

[3] There are not many Precambrian rocks that are able to preserve an original magnetization with adequate magnetic mineralogy for paleointensity determination. Data from Canadian mafic dykes and plutons were shown to be adequate recorders with consistent determinations of absolute paleointensity. Following an initial study [*Macouin et al.*, 2003] which was carried out on samples from 6 dykes swarms with ages between 2.4 Ga and 1 Ga, the present work focuses on the 1.2 Ga period in order to constrain further the possible existence of a “Precambrian dipole low” [*Macouin et al.*, 2004]. We conducted paleointensity experiments on samples of the Mackenzie giant radiating dyke swarm in the Canadian Slave Province [*Fahrig*, 1987]. These rocks are precisely dated by the U-Pb technique and have an age of 1267 ± 2 Ma [*LeCheminant and Heaman*, 1989]. Previous paleomagnetic studies (e.g., see summary by *Buchan and Halls* [1990]) indicated also that they are not characterized by large secondary components and established the primary origin of the characteristic remanent magnetization (ChRM).

2. Geological Setting and Sampling

[4] The Slave Province within the Canadian Shield is one of the oldest Archean cratons. The Mackenzie dyke swarm which intrudes this province is composed of a radial array of structures distributed fanned out over $\sim 100^\circ$ and extending over more than 2400 km [*Fahrig*, 1987] (Figure 1). This is actually one of the larger giant radial dyke swarms in the world. The focal region of the swarm in the northern Slave Province, contains coeval Coppermine flood basalts and the Muskox intrusion. The average dyke width is about 30 meters but dykes can attain a maximum width of 150 m. Petrography is characterized by medium to coarse gabbroic textures and mineralogy. *LeCheminant and Heaman* [1989] obtained a U/Pb baddeleyite age of 1267 ± 2 Ma for the swarm. According to the same authors, the entire dyke swarm was extruded over a short time span that did not exceed 5 Ma.

[5] Measurements have been conducted on 50 specimens involving 35 samples from 6 different dykes collected by one of us (R.E. along with W. R. A. Baragar and S. S. Gandhi) at different distances from the focus of the Mackenzie swarm: 400, 800 and 1000 km (Table 1, Figure 1) as part of a broader magnetic fabric and petrologic study of the swarm [*Ernst and Baragar*, 1992; *Baragar et al.*, 1996]. Sampling was concentrated within 3 meters of the finer-grained dyke margins. All samples were solar oriented.

3. Magnetic Mineralogy

[6] Rocks magnetic experiments were conducted at the Parc St Maur IPGP laboratory in order to investigate the suitability of the samples for paleointensity determinations. Weak-field thermomagnetic experiments (K-T) were performed on 6 samples using a KLY 2 and a KLY 3 Kappabridge system. There is no significant variation in low-field susceptibility before 550°C and then an abrupt drop of K over a narrow temperature interval (Figures 2a and 2b). The Curie points between 550 and 570°C are consistent with magnetite with a minor amount of titanium. Except for one sample (Figure 2c), the heating and cooling curves can be considered as reversible. The absence of major mineralogical changes during heating up to 610°C indicates that these samples are appropriate for paleointensity experiments. The low-temperature measurements are characterized by the presence of the Verwey transition at about -165°C (Figure 2d), which confirms that magnetite (or

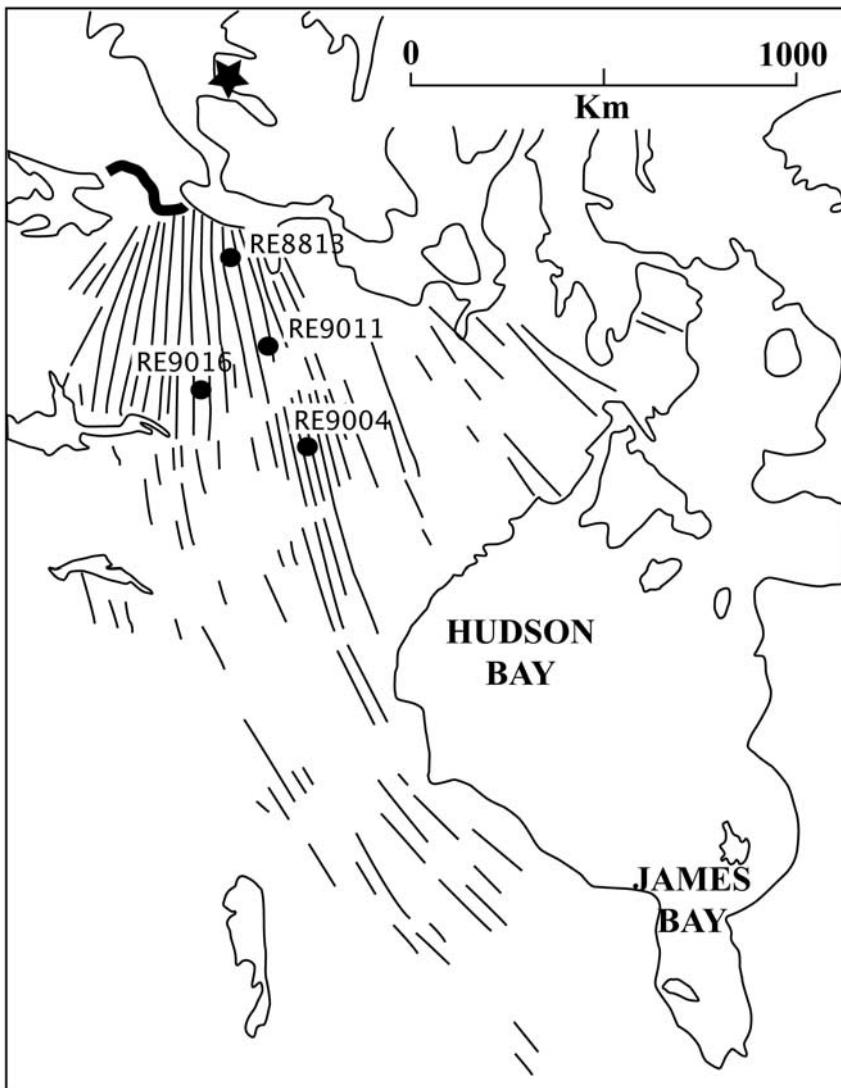


Figure 1. Simplified geological map of the Slave Province (Canada) and location of the Mackenzie radiating dyke swarm and sites (modified from Baragar *et al.* [1996]). The star marks the focal point of the swarm. The irregular black line marks the coeval Coppermine study.

magnetite with very minor amount of titanium or other components) dominates the remanence [Ozdemir *et al.*, 1993]. Alternatively, a similar behavior can also be attributed to non-stoichiometric (partially oxidized) magnetite [Dunlop and Ozdemir, 1997].

[7] Hysteresis cycles were performed on 5 fresh samples and 4 samples heated during Thellier experiments (several hours) with a vibrating magnetometer at the St Maur laboratory. The hysteresis parameters calculated after correcting for para- and diamagnetism [Day *et al.*, 1977] (Figure 3b) are consistent with pseudo-single domain (PSD) grain sizes. Hysteresis loops are not distorted (Figure 3a), which rules out the possibility of a mixture

between single domain and superparamagnetic grains or a combination of different magnetic grains [Tauxe *et al.*, 2002]. However, they probably represent also a mixture of monodomain and multidomain grains.

[8] These conclusions are consistent with the petrographic observations of Hoddych [1996] who observed intergrowth of nearly pure PSD magnetite with ilmenite lamellae in samples from several Precambrian Canadian dykes including the Mackenzie swarm. In contrast, microprobe analyses on Mackenzie dykes indicated that analyzed magnetites contain significant titanium with most results falling between Usp30 and Usp80 [see Baragar *et al.*, 1996, Figure 8]. Some exsolution lamellae

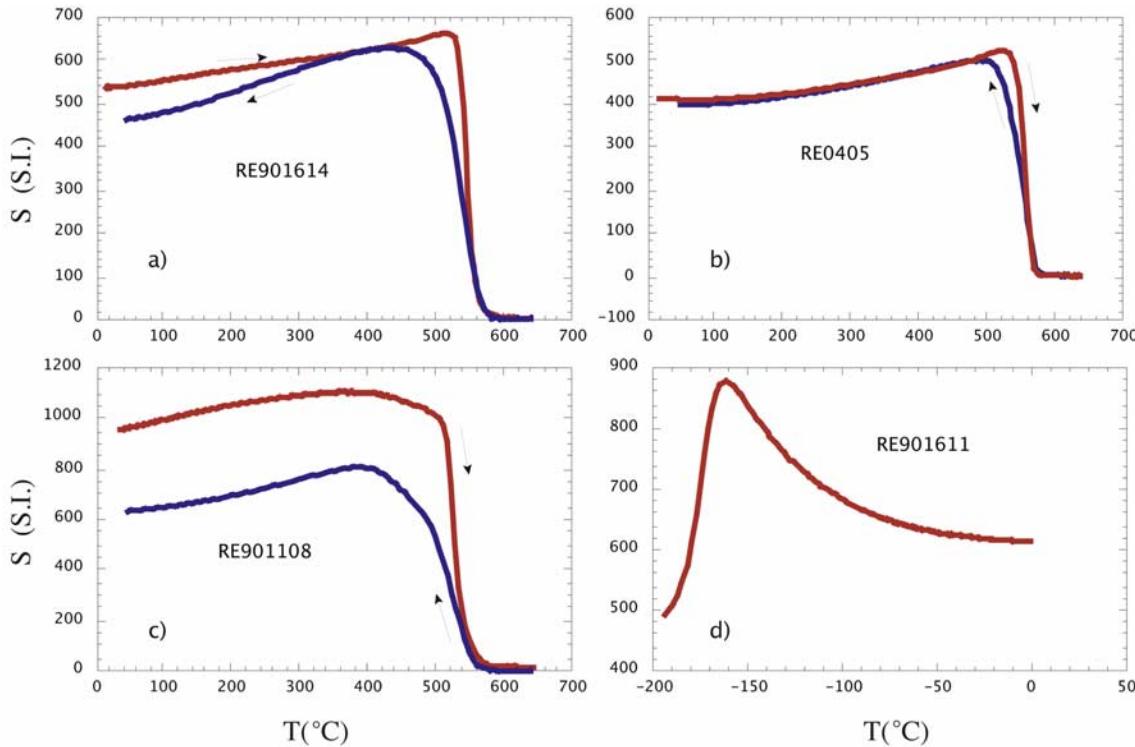
Table 1. Results of Paleointensity Determination^a

Sample	Site Location				Paleointensity								
	Lat.	Long.	D	I	H	\pm	Tmin	Tmax	N	q	w	f	Hlab
RE881304	67.12	244.28	209.5	-0.6	22.1	1.2	150	534	11	6.1	2	41.9	13
RE881307	67.12	244.28	223.1	22.6	16.4	0.4	525	549	12	23	7.2	54.5	22
RE8813, 400 km					19.3	4.0							
RE900405	62.76	249.62	216.2	32.3	9	0.5	470	529	7	5.7	2.5	44.6	13
RE900407	62.76	249.62	246.6	57.8	9.4	0.7	490	540	9	4	1.8	34.9	22
RE900411	62.76	249.62	227.7	62.4	9.5	1.4	536	544	5	1.5	0.9	35.4	22
RE9004, 1000 km					9.3	0.3							
RE901104	62.87	251.63	204.9	52.6	9.7	0.3	525	558	17	11	2.8	32.3	22
RE901106	62.87	251.64	268.2	18.5	6.5	1.5	300	490	4	1.2	0.9	48.7	13
RE901107	62.87	251.63	237.9	66.6	5.4	0.2	490	555	16	12	3.3	64.3	22
RE9011, 1000 km					7.2	2.2							
RE901611	64	248.78	253.1	23.5	6.6	0.4	508	544	7	5.2	2.3	59.3	13
RE901612	64	248.78	238.9	47.6	5.2	0.5	490	538	9	6.3	2.4	69.8	22
RE901612b	64	248.78	233	18.3	5.3	0.5	470	516	5	4.7	2.7	67.4	13
RE901614	64	248.78	234	32.9	5.9	0.5	516	544	6	5.5	2.7	54.6	13
RE901614b	64	248.78	235	35	6.5	0.4	400	535	5	5.3	3	51	22
RE9016, 800 km					5.9	0.7							

^a Column headings indicate the following: Sample is the sample number, Lat. and Long. are latitude and longitude of the site, D and I are the direction and the inclination, respectively, of the characteristic remanence determined in the same range of temperature as the paleointensity determination, H, the paleofield, \pm , the associated error, Tmin and Tmax, the interval of temperature used for the determination, N, number of points used for the determination, and q, w, and f represent the quality factors. Hlab is the field used during experiments. In bold are, from the left to right, site name, distance from focal point, and paleofield averaged by sites.

were observed, and such regions were avoided as much as possible during microprobe analysis. We conclude that the paleomagnetic remanence is not hosted in those magnetites that lack exsolution

lamellae and which yielded Ti-rich compositions. Instead, the magnetic experiments as well as microscopic observations converge to indicate that the remanent magnetization is carried by small

**Figure 2.** Examples of thermomagnetic curves obtained in weak fields. T and S represent the temperature and the susceptibility, respectively.

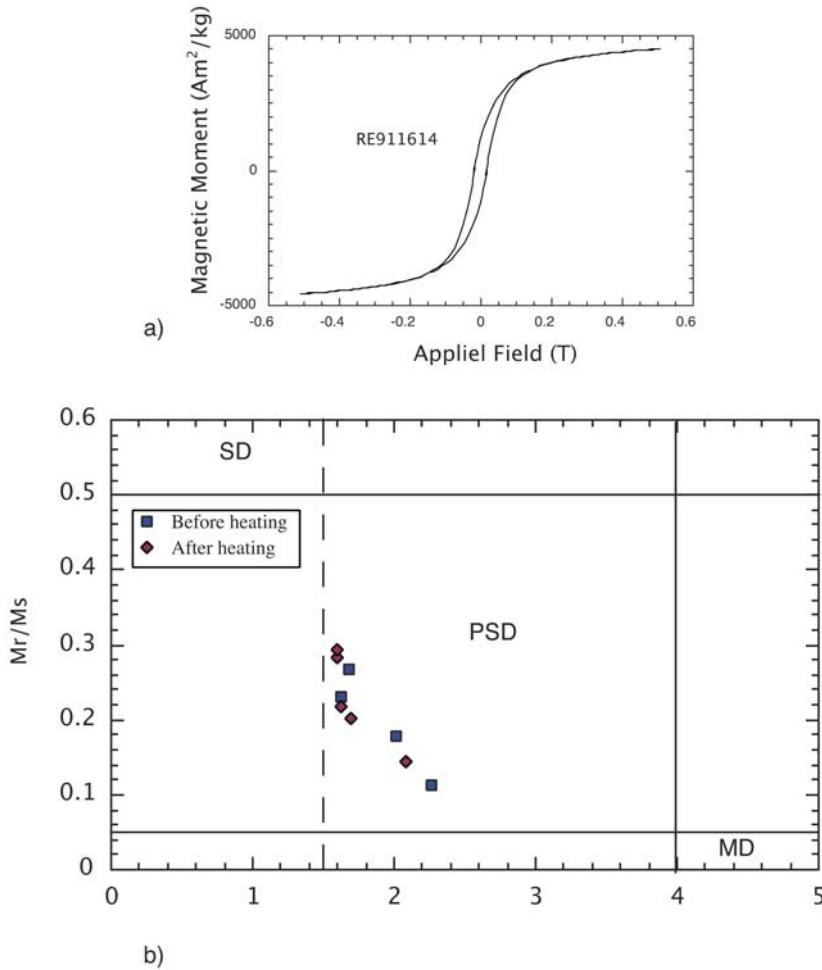


Figure 3. Hysteresis data. (a) Typical example of hysteresis curves (uncorrected) from small chip samples. (b) Hysteresis parameters from high-field cycles (diagram after Day *et al.* [1977]). Ratio of coercivity of remanence (H_{cr}) to coercivity (H_c) is plotted against the ratio of remanent magnetization (M_r) to saturation magnetization (M_s). Abbreviations are SD, single domain; PSD, pseudo-single domain; MD, multidomain.

domains of pure (or almost pure) magnetite separated by exsolution lamellae of ilmenite.

4. Full Vector Components

[9] The direction and intensity of magnetization were both derived from the experiments of absolute paleointensity. We used a double heating procedure [Coe, 1967; Thellier and Thellier, 1959] involving acquisition of pTRM followed by stepwise demagnetization in zero field at the same temperature. The unblocking temperatures of the NRM (Figure 4) were characterized by a very narrow distribution between 525 and 540°C. Consequently, the number of heating steps was reduced at lower temperatures since there was no change in magnetization and thus

no reason for increasing the possibility of mineralogical changes by accumulating heating time. Proper determination of the characteristic magnetization (ChRM) thus required temperature steps separated by only 2.5° within the range of temperature between 525°C and 540°C. Since such close temperature steps are of the same magnitude as typical longitudinal gradients found in most standard ovens, it was necessary to construct a new furnace [see Macouin *et al.*, 2003] with appropriate size to minimize the temperature gradient. This achievement allowed us control temperature steps to within 1.5°C and thus to perform repeatable measurements. Accurate temperature control was also essential to perform pTRM checks one step down from the last heating within the critical temperature range (mostly

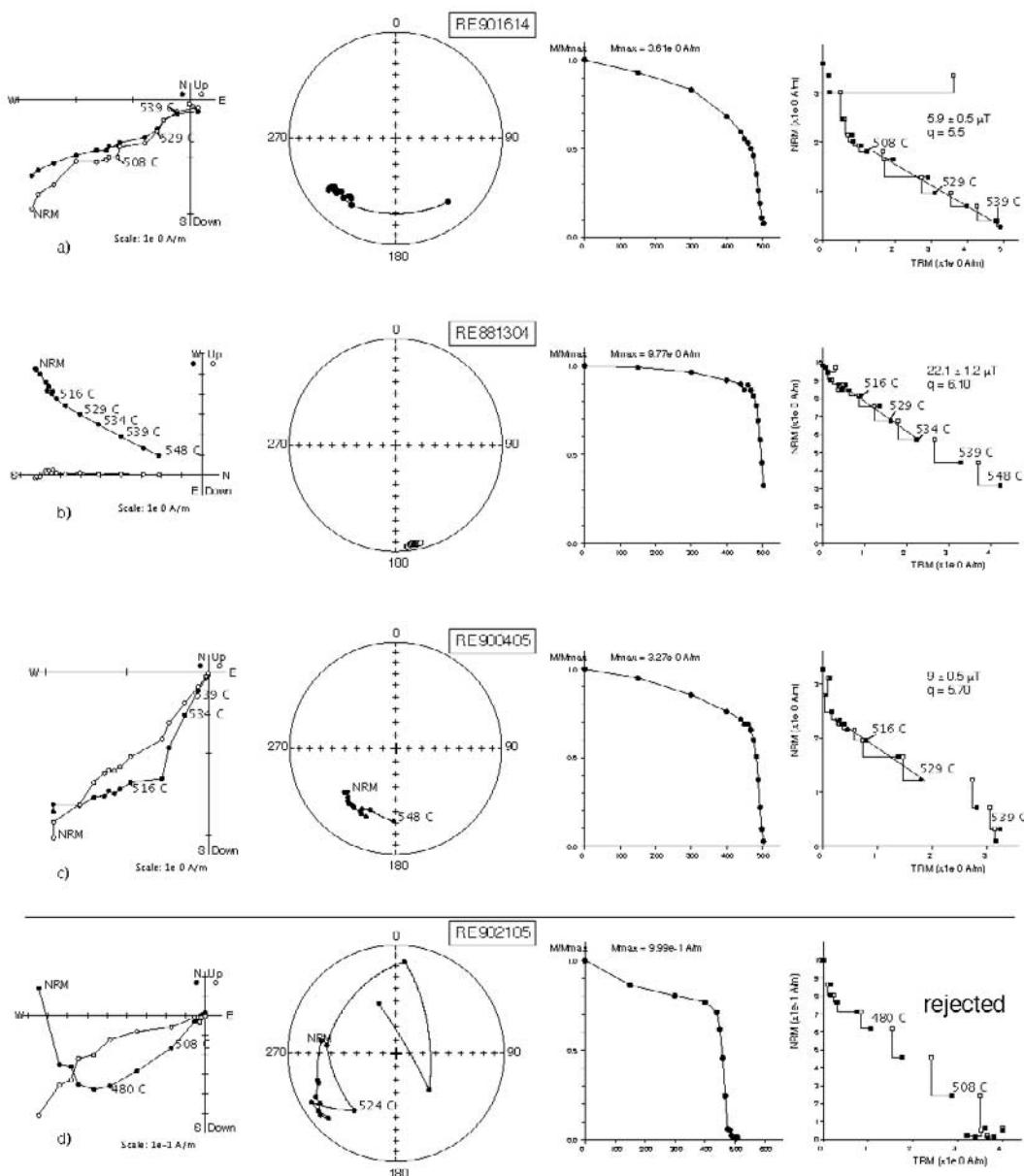


Figure 4. Examples of thermal demagnetization and intensity determinations. For each sample the following are shown: an orthogonal vector plot with solid (open) symbols for data projected onto the horizontal (vertical) plane, a stereographic projection, and a normalized plot of the intensity symbols for pTRM checks.

525°C–540°C over which the samples lost 50% to 80% of their initial magnetization. Other pTRM checks were performed two steps down ($T_{(i-3)}$) outside this critical interval.

[10] In Figure 4 are plotted some typical results incorporating the directional and intensity changes upon demagnetization as well as during partial remagnetization. The first characteristic is the presence of two components of magnetization in the initial NRM. The demagnetization diagrams of the

NRM are characterized by a secondary low-medium temperature component, which was not unblocked before 400°C and with a very weak moment compared to the high temperature component. It could be of viscous origin or be associated with a secondary mineral such as maghemite. However, there is no indication for conversion into hematite beyond 350°C. We are thus rather more inclined to favor a viscous origin that would be carried by grains with high T_{ub} . As indicated above, the first high temperature component was

isolated beyond 500°C and is characterized by a very narrow range of unblocking temperatures up to 540°C. Note that a similar distribution of T_{ub} was reported for other Proterozoic dykes of the Canadian Shield [Macouin et al., 2003] but over a slightly higher (20°C) range of temperatures. This distribution suggests that the high T_{ub} component is dominated by a very narrow distribution of monodomain grains of magnetite. Last, it is important to remember that the primary origin of the characteristic paleomagnetic direction (see summary and discussions by Fahrig and Jones [1969], Fahrig [1987], and Buchan and Halls [1990]) was previously established on the basis of baked contact tests [Irving et al., 1972]. The Arai diagrams, in which the NRM remaining at each temperature step was plotted against the TRM, can be classified within three categories. The first category (Figure 4a) includes all diagrams with linear NRM-TRM curves and positive pTRM checks over the same range of high temperatures as for the ChRM. The samples from the second category (Figures 4b and 4c) show linear TRM-NRM curves but negative pTRM checks at high temperatures. Occasionally, there is also dispersion at very high temperatures, which is evidently caused by additional mineralogical changes. The third category (Figure 4d) includes all the other samples with unsuccessful paleointensity experiments. The most common failure is the presence of negative pTRM checks at medium temperatures which reflect the occurrence of magneto-mineralogical evolution during the initial heatings. In most cases there is no linear relation between the NRM lost and the pTRM gained.

[11] The samples from the first two (successful) categories were accepted for determinations of absolute paleointensity. Our first selection criterion required the presence of a linear NRM-TRM segment over at least 30% of the total NRM. The presence of pTRM checks that do not deviate by more than 15% from the initial pTRM measurements was imposed as a second requirement. The choice of a limit of 15% was necessary to compensate for experimental difficulties, due to the abrupt decrease of the intensity over a very narrow range of temperatures, which enhanced the uncertainties in the measurements even with accurate temperature control.

[12] The early acquisition of the demagnetization can certainly be established on the basis of these experiments. However, it is much more delicate to determine whether the high temperature characteristic NRM component is a pure TRM, required for

paleointensity experiments, or a thermochemical remanent magnetization resulting from high temperatures oxidation below the Curie point of magnetite. In a recent paper, Smirnov and Tarduno [2005] suggest that this could be the case for the Canadian dykes with high unblocking temperatures. They claim that the TCRM/TRM ratio underestimates the true field value by a factor of four. These assumptions rely on two hypotheses. The first one is that this process indeed happens within the Canadian rocks, but so far we do not have any firm and conclusive indications. We note that low field determinations were obtained from various types of Precambrian rocks including other dykes with lower unblocking temperatures. Additional analyses are currently being performed to better constrain the origin of magnetization. It is obvious that additional data from other areas and various types of rocks are also crucial to answer this question. The second aspect relies on the TCRM/TRM ratio. The predictions for this ratio are very poorly constrained. Theoretical considerations [Stacey and Banerjee, 1974] predict that CRM should be lower than TCRM, but they were not tested by experimental data except at low temperatures [McClelland, 1996]. The unique experimental approach [Stokking and Tauxe, 1990] was done for single domain hematite and goethite and is thus of no direct concern for the present study. Thus we do not discard this hypothesis but consider that it relies on few experimental evidences and observations. In the present state we simply face it with the large number of experimental work that constructed the existing data set.

5. Results

[13] The individual ChRM directions of the thirteen samples (that passed the above criteria) were obtained by least squares fitting of the individual vectors through the origin of the demagnetization diagrams. Since they were obtained at various latitudes, we calculated the corresponding Virtual Geomagnetic Poles (VGPs) and their averaged values (Figure 5), which were found in good agreement with the mean pole position (4°N, 190°E and $\alpha_{95} = 5^\circ$) for the 1.27 Ga Mackenzie dykes swarm [Buchan and Halls, 1990; Irving et al., 1972].

[14] Thirteen successful determinations of paleointensity (Table 1) were obtained from four dykes, which represents a success rate of 30%. The paleo-field values range between 4.3 and 22.1 μT . Three dykes (RE9004, RE9011 and RE9016) are charac-

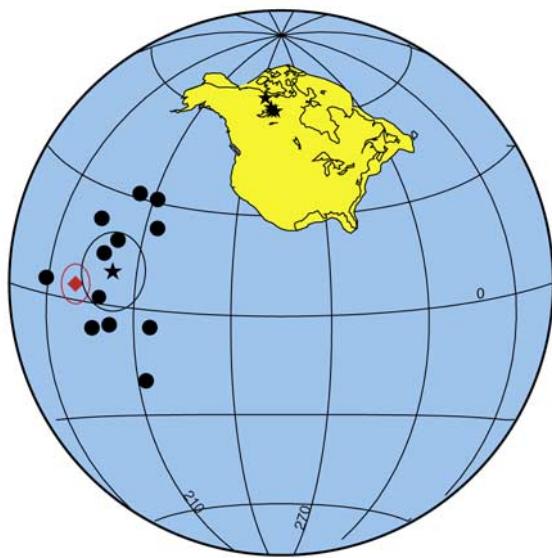


Figure 5. Stereographic projection of paleomagnetic poles calculated from individual directions from successful paleointensity samples. The star represents the mean pole derived from the thirteen individual directions. The diamond marks the expected paleomagnetic pole [Buchan and Halls, 1990]. Small stars represent the site locations.

terized by very similar results with an average field intensity of $6.6 \pm 0.96 \mu\text{T}$. In contrast, the mean paleointensity of $19.25 \pm 4.03 \mu\text{T}$ obtained for the RE8813 dyke is significantly higher.

[15] Virtual Dipole Moments (VDM) are commonly used to compare paleointensity results acquired at different paleolatitudes. The underlying assumption is that the Precambrian magnetic field was dominated by a geocentric axial dipole. Following a precursory analysis by Evans [1976], Kent and Smeethurst [1998] reanalyzed the global paleomagnetic database and proposed that the Precambrian field was characterized by significant octupole (25%) and quadrupole (10%) contributions. Recently, McFadden [2004] demonstrated that a period of at least 5000 Myr is required for this test to be effective. Controversial arguments can be brought up on several aspects including the stability of the results within various time intervals. The 1100 Ma old Keweenawan flood basalts from the Mid-Continent rift of North America exhibit an apparent 15–20° asymmetry between normal and reversed polarity data along the younger leg of the Logan Loop [e.g., Pesonen and Nevanlinna, 1981; Pesonen and Halls, 1984]. However, inclusion of paleomagnetic results from related carbonatites [Symons et al., 1994] and reassessment of data from the Mamainse Point volcanics [e.g., Ernst and

Buchan, 1993; Schmidt and Williams, 2003] suggests that polar wander is more likely the explanation. Furthermore, the circa 1200 Ma Strathcona Sound Formation in the Upper Borden Basin sequence in the Bylot Basin of northern Canada exhibits 6 symmetric reversals [Fahrig et al., 1981; Ernst and Buchan, 1993]. Gallet et al. [2000] reported antipodal directions for about 15 geomagnetic reversals recorded in two coeval sedimentary formations in Siberia that were deposited between 1050 and 1100 Ma. These results as well as other records of reversals from older rocks (2.45 Ga [Halls, 1991]) with symmetrical reverse and normal directions suggest that there is no strong evidence or indication that the field was not already dominantly dipolar. The large latitudinal distribution of the Mackenzie dyke swarm could provide a powerful test for the magnetic field polarity asymmetry but the sites were counterclockwise rotated by about 90° with respect to the present orientation of the Canadian craton (Figure 5) so that their paleolatitudes do not differ by more than 1.5° (between 16° and 17.5°).

[16] The mean VDMs derived from the dykes RE9004, RE9011 and RE9016 are between 1.3 ± 0.2 and $1.8 \pm 0.3 \times 10^{22} \text{ Am}^2$ and slightly above ($4.5 \pm 0.9 \times 10^{22} \text{ Am}^2$) for the two samples from RE8813. Site RE8813 is located 400 km away from the focus point of the dyke swarm while the other ones lie 800 to 1200 km away. The difference between these mean VDMs can be caused by secular variation in the mean dipole field or by large non dipole components. Baragar et al. [1996] discussed petrographic differences between sites and suggested that the 400 km dykes were associated with shallower intrusion depths. This results in a small difference between the age of these 400 km-distance dykes and those that were emplaced further away at deeper levels. Thus the difference between these VDMs simply reflects a variation of the Earth's magnetic field strength during the 5 Ma period which was taken by the emplacement of the dyke swarm [LeCheminant and Heaman, 1989].

6. Integration Within the Existing Database and Discussion

[17] The present data set must be integrated with previous results published for the same period. Following the same criteria as for this study, we extracted the data points obtained by double heating techniques which included pTRM checks (referred as T+) from the PINT 03 database [Perrin and Schnepp, 2004; Perrin et al., 1998]. In their

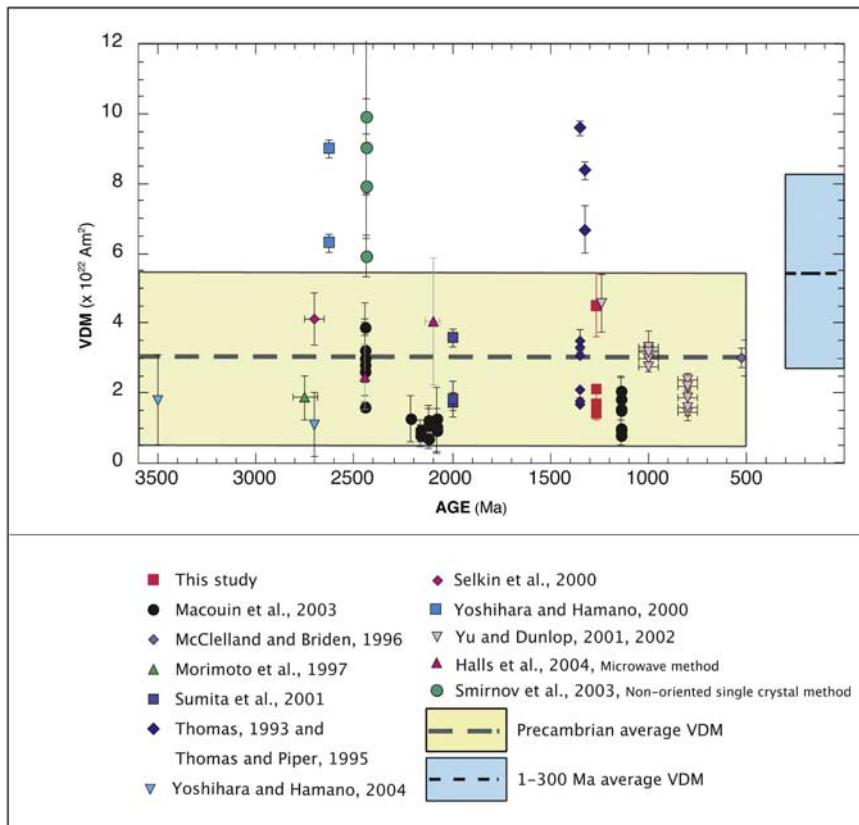


Figure 6. Variation of Virtual Dipole Moment (VDM) over the 3–0.5 Ga period.

recent compilation, *Dunlop and Yu* [2004] build up three categories, the first one requiring positive backed contact tests, partial TRM checks and 10 or more determinations. Interestingly the results did not change strikingly between the A category and the entire database. As mentioned before the mean VDMs remain very similar ($3 \pm 1 \times 10^{22} \text{ Am}^2$ and $3.21 \pm 2.2 \times 10^{22} \text{ Am}^2$, respectively) but in the absence of a Gaussian distribution. This is in contrast with the situation reported for the 0.3–5 Ma interval for which the results of the Thellier experiments are different from those obtained with other techniques [*Selkin and Tauxe*, 2000].

[18] The successive VDMs resulting from our selection are plotted in Figure 6 as a function of age. The time interval surrounding the ~1 Ga period is one of the best documented Precambrian period in terms of paleointensity. However, all studies were obtained from Canadian sites, which reflects difficulties in finding appropriate rocks of this age from other old cratons. Two exceptions [*Yoshihara and Hamano*, 2004; *Sumita et al.*, 2001] were recently published from Africa and Western Australia, respectively, which indicate a low field intensity during the Archean but these results must be taken with some caution since they

were obtained from overprints of chemical or thermal origin. Most records were obtained in the Superior province of the Canadian Shield [*Macouin et al.*, 2003; *Yu and Dunlop*, 2001, 2002]. The 1240 Ma Tudor Gabbro, the 1.141 Ga Abitibi dyke swarm and the ~1 Ga Cordova Gabbro yield VDMs between 5 and $1.1 \times 10^{22} \text{ Am}^2$. Three data points obtained from 1.3 Ga old basalts in Greenland [*Thomas and Piper*, 1995; *Thomas*, 1993] contrast with the above results with VDMs between 6 and $10 \times 10^{22} \text{ Am}^2$. Prior to this time, there is a 700 Ma long period (between 1.3 and 2 Ga) which lacks paleointensity data.

[19] Prior to 2 Ga, more than twenty analyses [*Halls et al.*, 2004; *Yoshihara and Hamano*, 2004; *Macouin et al.*, 2003; *Smirnov et al.*, 2003; *Sumita et al.*, 2001; *Selkin et al.*, 2000; *Morimoto et al.*, 1997] satisfying our criteria are consistent with low VDMs (Figure 6). Two studies provide VADM values greater than $5 \times 10^{22} \text{ Am}^2$ [*Smirnov et al.*, 2003; *Yoshihara and Hamano*, 2000] and there is a concentration of VDMs lower than $2 \times 10^{22} \text{ Am}^2$. Of particular interest are the VDMs obtained using recently developed techniques or selection of materials. The first field value was recorded by the Burakovka dikes from separated

plagioclase crystals [Smirnov *et al.*, 2003] with the same age as the low VDM of the Matachewan dykes [Macouin *et al.*, 2003]. Two studies [Halls *et al.*, 2004; McArdle *et al.*, 2004] were recently performed using microwave experiments. Interestingly the first one involves the 2.45 Ga Matachewan dykes and indicates a mean VDM of $2.5 \pm 0.9 \times 10^{22}$ Am 2 which is in excellent agreement with the results of Macouin *et al.* [2004] ($2.8 \pm 0.9 \times 10^{22}$ Am 2). Halls *et al.* [2004] and Dunlop and Yu [2004] emphasize that these results disagree with the $8.4 \pm 2.1 \times 10^{22}$ Am 2 VDM obtained by Smirnov *et al.* [2003]. This difference evidently is too large to be explained by secular variation. Given the uncertainties on the age estimates, they could simply reflect the dispersion of the field. We reiterate that the database is too small to draw any firm conclusion but also that the distribution is not a Gaussian distribution but rather shows two peaks, a large one with small values and a second and smaller one with values closer to the present field. This reinforces the importance of additional results outside the Canadian Shield but also the necessity of comparing various techniques using the same samples. The second microwave experiments [Halls *et al.*, 2004; McArdle *et al.*, 2004] concerned the 2100 Ma dykes from the Biscotasing, Marathon, and Fort Frances swarms. The mean VDM of $4.1 \pm 1.8 \times 10^{22}$ Am 2 does not overlap with the value of $1 \pm 0.1 \times 10^{22}$ Am 2 derived from the initial study by Macouin *et al.* [2003] using the classical Thellier technique. We are inclined to link this difference with the experimental procedures, given the very different character of the demagnetization diagrams obtained by each technique. It is clear that additional experiments will help clarify the debate on this matter.

[20] The mean VDM obtained by averaging the mean values for each cooling unit (per dyke or per site for lava flows and plutons) with ages between 800 Ma and 3500 Ma is $3.1 \pm 2.5 \times 10^{22}$ Am 2 . Unfortunately, the absence of data within the 800 and 300 Ma interval precludes any comparison with this period. Prior to 300 Ma, the mean VDM derived from the selected records of the PINT 03 database is $5.4 \pm 3 \times 10^{22}$ Am 2 for the 1–300 Ma interval [Macouin *et al.*, 2004], thus almost twice higher than the mean Precambrian field. These two estimates are characterized by a large variance. The distributions of the VDMs are significantly different for these two periods [Macouin *et al.*, 2004], which may either be caused by an insufficient number of data for the Precambrian or may indicate that the two distributions are

indeed different and thus reveal different mechanisms associated with the field regeneration. Although less attractive, the present number of records favors the first hypothesis.

7. Conclusion

[21] We herein investigated samples from the 1270 Ma Mackenzie dyke swarm in the Slave province (Canada). The magnetic mineralogy associated with the primary ChRM magnetization is relatively simple, consisting largely of single or pseudo-single domain grains of magnetite or low-titanium titanomagnetite. This mineralogy and the distribution of the unblocking temperatures suggest that the characteristic remanence is a primary thermoremanent magnetization.

[22] Thirteen successful paleointensity determinations were obtained from four dykes using a special furnace to demagnetize the samples characterized by a narrow range of unblocking temperatures and to allow temperature steps as small as 1.5°C. The VDMs range between 1.3 ± 0.2 and $4.5 \pm 0.9 \times 10^{22}$ Am 2 .

[23] The present results, combined with other reliable data from the 3.5–1.0 Ga period do not support the model of a dramatic rise in paleointensity in the 2.7–2.1 Ga period [Hale, 1987]. In the present state of the database, there is a clear dominance of low field intensity with a mean value of about $3.1 \pm 2.5 \times 10^{22}$ Am 2 during the Precambrian. However, the existence of a few low values sometimes coexisting within the same short time intervals remains to be elucidated. Because some were obtained using new approaches, it becomes important to test the validity of the determinations by using multiple techniques.

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