JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 106, NO. B11, PAGES 26,331–26,343, NOVEMBER 10, 2001

Paleointensity determination on the Late Precambrian Tudor Gabbro, Ontario

Yongjae Yu and David J. Dunlop

Geophysics, Department of Physics, University of Toronto, Toronto, Canada

Abstract. Paleointensity determination by the Coe-modified Thellier method was carried out on the ≈1100 Ma Tudor Gabbro of southern Ontario. The Tudor has a bivectorial natural remanent magnetization (NRM). The C NRM is a Grenvillian uplift thermoviscous remagnetization which overprints the A NRM, the surviving primary thermoremanent magnetization (TRM) of the Tudor intrusion. A sharp junction at ≈530°C between the A and C vectors in laboratory thermal demagnetization matches the peak metamorphic reheating temperature of 480-500°C in nature when the difference in time scales is taken into account. The unblocking temperature distributions of the A NRM and of laboratory TRM also match. A is carried by single-domain or small pseudo-single-domain grains, probably magnetite rods exsolved in plagioclase or pyroxene at temperatures above 580°C. Arai plots of NRM versus partial TRM are linear from 520°C or 530°C to 580°C for 45 specimens of 19 samples from 9 sites. The 45 reliable paleointensity values from these specimens give a virtual axial dipole moment (VADM) of $4.6 \pm 0.8 \times 10^{22}$ A m^2 around 1100 Ma. This is about one half of the modern field intensity but is similar to the mean Cretaceous-Cenozoic intensity and to most Archean and Proterozoic VADMs. Keweenawan rocks, whose ages are within ≈20 Myr of the Tudor Gabbro, have a VADM of 11.4×10^{22} A m^2. The difference between the Tudor and Keweenawan VADMs is similar to differences recorded over similar time intervals in the Cretaceous and seems to represent normal variation in geomagnetic field strength throughout Earth's history.

1. Introduction

The record of geomagnetic field intensity over geological time is a first-order constraint on the operation of the core geodynamo and possible growth of the inner core. During Precambrian time, competing models of thermal evolution of the core predict distinctively different variations of Earth's magnetic dipole moment, but reliable Precambrian paleointensity data are presently too few to test these models and predictions [*Prévot and Perrin*, 1992]. To date, only 12 Thellier-type paleointensity determinations have been reported for Precambrian rocks of all ages from all continents (Table 1). If stringent acceptance criteria typical of modern work are applied (e.g., see section 6), the number of Precambrian results is reduced still further. For example, only six of the paleointensities tabulated were obtained using more than five specimens and the older studies lack partial thermoremanent magnetization (pTRM) checks, now considered indispensable.

In this paper, we report reliable Thellier paleointensity determinations for 45 specimens (19 samples) of the late Precambrian Tudor Gabbro of southern Ontario. The Tudor Gabbro is unique among formations from the Grenville Province of the Canadian Precambrian Shield in its relatively low metamorphic grade, its preservation of a thermoremanent magnetization (TRM) dating from initial cooling, and the dating of its TRM within narrow limits by ^{40}Ar/^{39}Ar geochronology. Section 2 reviews this evidence.

Paper number 2001JB000213.
0148-0227/01/2001JB000213$09.00

2. Geology, Paleomagnetism, and Geochronology

The Tudor Gabbro lies in the Hastings Lowlands (the Elsevir terrane of *Easton* [1992]) in the Central Metasedimentary Belt of the Ontario Grenville Province (Figure 1). The Elsevir is outlined by the kyanite-sillimanite isograd [*Anovitz and Essene*, 1990, Figure 2] and represents a "window" of low metamorphic grade in the otherwise high-grade Grenville Province. The Tudor Gabbro is in the lowest-grade core of the Elsevir, below both the oligoclase and zoned albite isograds [*Lumbers*, 1969]. On the basis of compatible garnet-pyroxene geobarometers and geothermometers, estimated metamorphic pressures in the core region reached 4-5 kbar, and temperatures were 480-500°C [*Anovitz and Essene*, 1990, Figures 3 and 6]. Thus the Tudor Gabbro and its surroundings were buried to (by Grenvillian standards) shallow (≈15 km) depths and were never reheated above ≈500°C in the Grenvillian orogeny.

The Tudor Gabbro has preserved a recognizable contact metamorphic aureole dating from its intrusion, although rocks from the contact zone (unlike the gabbro itself) were totally overprinted magnetically in the Grenvillian orogeny and are unusable paleomagnetically [*Dunlop et al.*, 1985]. Postkinematic intrusions like the Tudor, Cordova, and Thanet gabbros (Figure 1) were all emplaced between 1200 and 1100 Ma, on the basis of U/Pb and Ar/Ar dating. The Thanet and Cordova Gabbros have hornblende ^{40}Ar/^{39}Ar ages as old as 1200 and 1150 Ma, respectively [*Berger and York*, 1981; *Lopez-Martinez and York*, 1983], but neither body preserves a primary TRM [*Buchan*, 1978; *Dunlop and Stirling*, 1985]. The characteristic (highest unblocking temperature/coercivity) component of natural remanent magnetization (NRM) in both is an A magnetization dating from ≈950 Ma [*Berger et al.*, 1979;

Table 1. Previous Precambrian Thellier-Type Paleointensity Determinations [a]

Formation or Lithology	Paleointensity			Dating		
	VDM, $\times 10^{22}$ A m^2	N	Ref.	Age, Ma	Methods	Ref.
Komati Volcanics	4.69 ± 0.13	4	1	3470 ± 20	$^{40}Ar/^{39}Ar$	10
Abitibi metaperidotites	5.00 ± 2.54	4	2	2765 ± 42	U/Pb, Sm/Nd	11,12
Dolerite dike (west Greenland)	1.9 ± 0.6	12	3	2752 ± 63	K/Ar	3
Abitibi gabbro/basalt	3.22 ± 1.61	4	2	2703 ± 2	U/Pb	11
Stillwater Complex	4.1 ± 0.5	11	4	2703 ± 50	U/Pb, Sm/Nd	13
Diabase dikes (Dwyer Lake)	6.3 ± 0.2	15	5	2631 ± 11	U/Pb	14,15
Diabase dikes (Yellowknife)	9.0 ± 0.2	15	5	2631 ± 11	U/Pb	14,15
Matachewan dikes	5.83 ± 0.51	4	2	2452 ± 3	U/Pb	16
Preissac dikes	18.26 ± 5.50	3	6	2150 ± 20	$^{40}Ar/^{39}Ar$	17
Sudbury Norite	3.90 ± 0.97	52	7	1850 ± 1	U/Pb	18
Keweenawan rocks	11.4 ± 3.6	54	8	1092 ± 5^{b}	U/Pb	19
Gananoque Diabase	5.76 ± 1.39	4	9	817 ± 70	K/Ar	20

[a] N is the number of specimens averaged; Ref. are the references: 1, *Hale and Dunlop* [1984]; 2, *Hale* [1985]; 3, *Morimoto et al.* [1997]; 4, *Selkin et al.* [2000]; 5, *Yoshihara and Hamano* [2000]; 6, *Hale* [1987]; 7, *Schwarz and Symons* [1970]; 8, *Pesonen and Halls* [1983]; 9, *Schwarz and Symons* [1969]; 10, *Lopez-Martinez and York* [1983]; 11, *Nunes and Jensen* [1980]; 12, *Zindler et al.* [1978]; 13, *Premo et al.* [1990]; 14, *Henderson et al.* [1987]; 15, *Dudas et al.* [1990]; 16, *Heaman* [1995]; 17, *Hanes and York* [1979]; 18, *Krogh et al.* [1984]; 19, *Palmer and Davis* [1987]; 20, *Wanless et al.* [1966].

[b] Two U/Pb dates [*Palmer and Davis*, 1987] for Keweenawan reversed rocks. Thirty-four of fifty-four reliable paleointensities [*Pesonen and Halls*, 1983] were for reversed polarity samples.

Cosca et al., 1991], and the lower unblocking temperature B component of NRM is dated by $^{40}Ar/^{39}Ar$ on feldspars around 800 Ma [*Berger et al*, 1979; *Dallmeyer and Sutter*, 1980]. The A→B succession defines a Grenville Track in the Laurentian apparent polar wander path [*Costanzo-Alvarez and Dunlop*, 1998; *Weil et al.*, 1998] that records postmetamorphic uplift and cooling from ≈500°C to ≈200°C between 950 and 800 Ma.

Hornblendes from the Tudor Gabbro define an 1110 Ma isochron [*Baksi*, 1982], recording the time of cooling through the Ar closure temperature of 590°C. Since the Tudor has never been reheated above 500°C, 1110 Ma is the age of primary cooling of the intrusion. Like the Thanet and Cordova Gabbros, the Tudor has two components of NRM, called A and C by *Dunlop et al.* [1985]. However, the A paleopole does not fall on the Grenville Track. The A NRM cannot postdate the Grenville Track because its high unblocking temperatures and coercivities (see below) are not those of a very low temperature (<200°C) late post metamorphic overprint. Rather, they are appropriate to a high-temperature (>500°C) primary TRM that predates burial reheating and was only partially overprinted by it (producing the C component). Thus the A NRM predates the Grenville Track .

This interpretation implies a convergent junction between Grenvillian and Interior Laurentian polar wander tracks and a scissor-like convergence of these plates around 1050 Ma [*Dunlop et al.*, 1985]. Docking of a terrane to the southeast is recorded by 1024-1061 Ma (mean 1041 Ma) U/Pb ages of monazites in the Bancroft terrane, immediately northwest of the Elsevir [*Mezger et al.*, 1991], and a similar style of convergence has been proposed for accretion of other terranes in Grenvillia [*Culshaw et al.*, 1994]. The divergence recorded paleomagnetically by the A NRM of the Tudor Gabbro between the Elsevir terrane and Interior Laurentia and the subsequent ocean closing recorded by the ≈1050 Ma convergence of polar wander paths (and by ≈1040 Ma U/Pb monazite ages) form part of the Grenvillian orogenic cycle, a chain of similar events occurring at different times in the interval 1160-970 Ma (U/Pb on zircon, monazite and titanite) [*Rivers et al.*, 1989]. The fact that the Tudor paleomagnetic record is compatible with overall Grenvillian tectonic history strengthens our confidence that the Tudor A magnetization is not anomalous in direction, nor a late postmetamorphic chemical remanent magnetization (CRM), but a primary record of initial cooling of the intrusion.

3. Sampling and Basic Paleomagnetic Properties

Figure 2 is geological map of the Tudor Gabbro, based on the work by *Lumbers* [1969]. The Tudor Volcanics, Glanmire

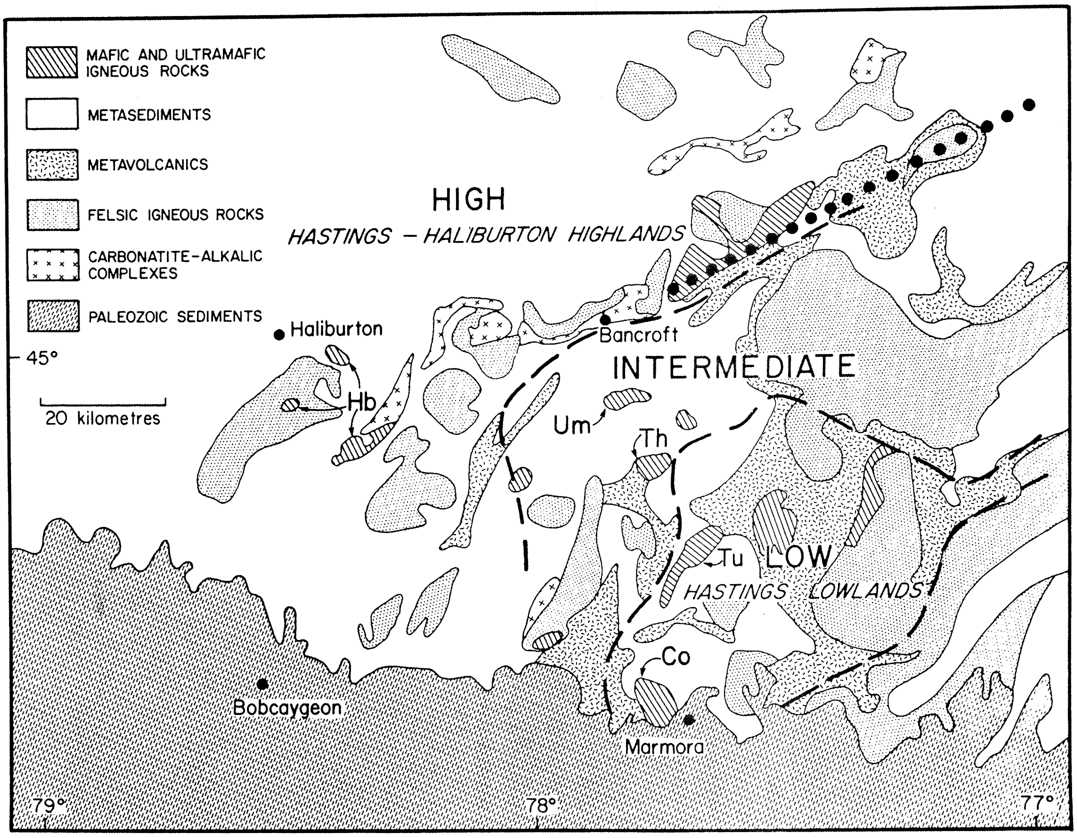

Figure 1. Metamorphic zonations in part of the Central Metasedimentary Belt, Grenville Province, southern Ontario. The low- and intermediate-grade zones fall within the Elsevir terrane of *Easton* [1992], while the high-grade Bancroft terrane is to the northwest. The Tudor, Thanet, and Cordova Gabbros are designated by Tu, Th, and Co.

Complex, and Hastings and Glanmire Marbles, with U/Pb ages of 1300-1250 Ma [*Lumbers*, 1969], were intruded by the Tudor Gabbro at 1100 ± 20 Ma [*Baksi*, 1982]. We concentrated our sampling in deep roadcuts along Highway 62, where fresh material is exposed. Samples from nine of these sites (numbered in Figure 2) gave paleointensity results of acceptable quality. Samples from most other sites (open squares in Figure 2) had weak and/or unstable NRMs.

The Tudor Gabbro has preserved its contact metamorphic aureole. The best test of primary TRM would be a positive contact test on the baked contact rocks in this aureole and beyond it into unbaked country rocks. To this end, three traverses were sampled by *Dunlop et al.* [1985] to the northeast, northwest, and south of the Tudor Gabbro. All Glanmire and Hastings Marble samples were very weakly magnetized and had no recognizable A or other stable NRM. Some Tudor Volcanics samples did exhibit swings of the NRM vectors toward an A-like direction similar to the characteristic NRM of the Tudor Gabbro (Figure 3), but these samples did not preserve a consistent underlying primary NRM of their own. The contact test was therefore indeterminate.

In contrast, the Tudor Gabbro samples from the nine successful sites were stably and consistently magnetized, as well as being very fresh in appearance. The thermal demagnetization behavior shown in Figure 3 is typical of all samples (compare Figures 5a-5d). A magnetization with north to NE declination and shallow inclination, the C component, is demagnetized from 20°C to 520°C or 530°C. From 520-530°C to 580°C, the Curie point of magnetite, the A component, with NW declination and steep negative inclination, is demagnetized. The junction between C and A on vector projections is quite sharp. Because A and C are almost perpendicular vectors, the total intensity decay curve is almost flat below 520°C, even though C typically represents about one third of the NRM.

The sharp junction between the C and A vectors in thermal demagnetization implies that C is a thermal or thermoviscous overprint of A [*Dunlop*, 1979; *Dunlop et al.*, 1997a, 1997b]. Post-Grenvillian uplift and cooling were very slow, cooling from ≈550°C to ≈300°C having taken ~150 Myr [*Berger et al.*, 1979; *Dallmeyer and Sutter*, 1980; *Cosca et al.*, 1991, 1992]. As a result, the maximum burial reheating temperature in nature of 480-500°C [*Annovitz and Essene*, 1990] (see also section 2) corresponds, for single-domain magnetite (see section 4), to a laboratory demagnetization temperature of 520-540°C on a timescale of minutes to hours [*Pullaiah et al.*, 1975; *Dunlop et al.*, 2000]. This is exactly the range observed in our thermal demagnetization experiments. In other words, the laboratory behavior of the C NRM agrees in every respect with that of a thermal overprint stabilized by cooling from a maximum burial temperature of ≈500°C during uplift following Grenvillian metamorphism.

What of the A magnetization? It is not a Grenvillian thermal overprint because its unblocking temperatures are too high. It could be a CRM, which can produce unblocking temperatures much above the blocking temperature at which chemical change occurred. However, its direction does not match A or B poles from the Grenville Track, and so it would have to be a late overprint acquired at temperatures below 200°C. In that case, it is surely curious that it has generated unblocking temperatures exclusively above 520-530°C and none below. The match

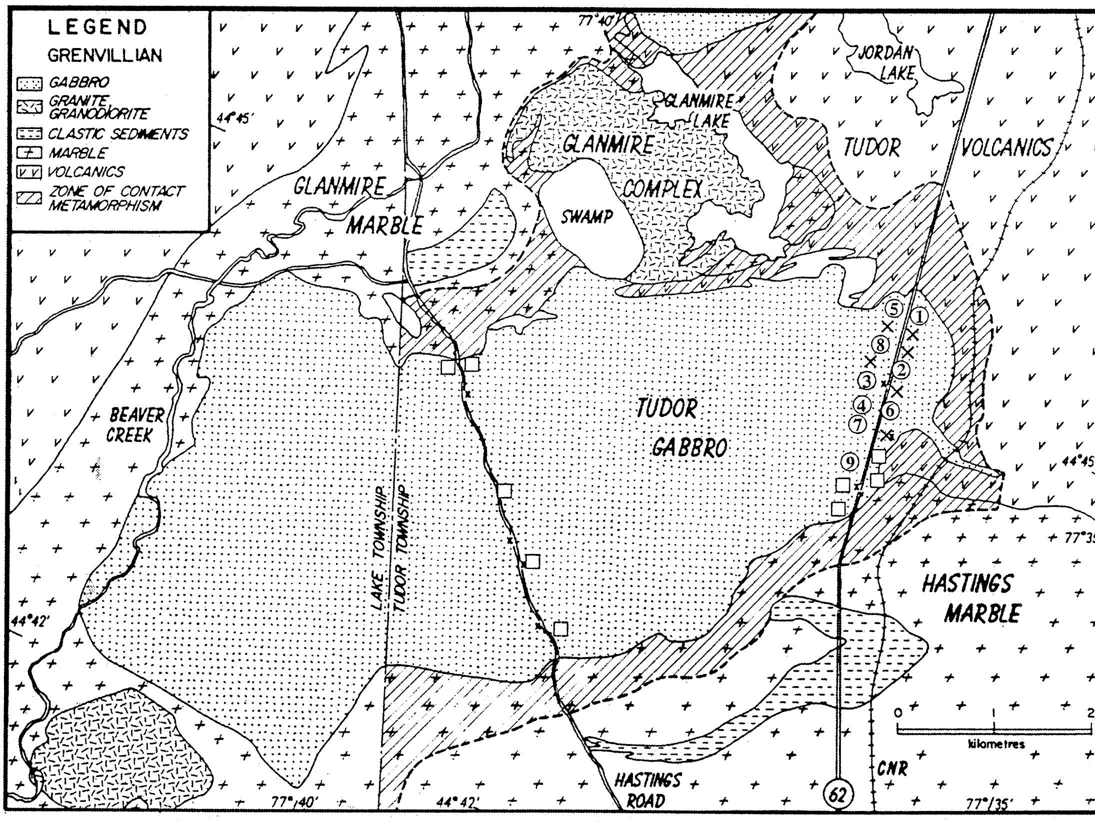

Figure 2. A geological map of the Tudor Gabbro and the formations it intrudes (Glanmire and Hastings Marbles and Tudor Volcanics). The limit of the contact metamorphic aureole around the Tudor Gabbro (mapped in Tudor Township only) is shown by the dashed line. Successful sites are shown by numbered circles, and unsuccessful sites are shown by crosses and squares.

between this laboratory unblocking temperature limit and the maximum burial reheating temperature would then have to be coincidental, since A would overprint C, not vice versa. Furthermore, the sharp junctions between A and C on vector plots would also be inexplicable, since CRM overprints generate rounded rather than sharp junctions [*Dunlop*, 1979; *McClelland-Brown*, 1982].

Could A be an early, rather than a late, CRM? Although possible, this is also unlikely. Figure 4 illustrates typical alternating field (AF) demagnetization curves of NRM and anhysteretic remanent magnetization (ARM) for samples in which A dominates the NRM. The initial plateau of no demagnetization below 10-15 mT, the strong concentration of high coercivities (30-80 mT), and the inflected sigmoid shape of the curves are all hallmarks of single-domain magnetite [*Dunlop and Özdemir*, 1997, Chapter 8]. The source is likely submicroscopic magnetite needles and rods exsolved in plagioclase and/or pyroxene, which have been much studied in connection with the stable NRM of gabbros and diabases [*Hargraves and Young*, 1969; *Evans and Wayman*, 1970, 1974; *Davis*, 1981; *Dunlop and Prévot*, 1982; *Zhang and Halls*, 1995]. These magnetite inclusions exsolve from their host minerals during slow cooling at temperatures above the Curie point of 580°C [*Garrison and Taylor*, 1981]. Their NRM is therefore a TRM rather than a CRM.

Summarizing, there are four lines of evidence that favor a primary TRM origin for A.

1. The A NRM has the AF and thermal demagnetization characteristics of a remanence carried by single-domain magnetite exsolved in silicate host minerals. Such exsolution occurs above the magnetite Curie temperature, so that A must be a TRM, not a CRM.
2. Garnet-pyroxene geothermometers indicate a maximum burial reheating temperature of 480-500°C for the Tudor area. The C component of NRM has maximum laboratory unblocking temperatures of 520-530°C, which, allowing for the difference in cooling rates, match the burial temperatures closely. Junctions between A and C vectors in the course of thermal demagnetization are sharp, indicating that C is a thermoviscous overprint of A. Therefore A is older than C and predates Grenvillian burial and uplift.
3. Tectonic interpretations made on the assumption that A predates docking of the Elsevir terrane are consistent with the overall pattern of Grenvillian orogenic events.
4. Some Tudor Volcanics samples within 1-2 km of the intrusive contact have an NRM component similar to the Tudor Gabbro A. However, an underlying stable pre-1100 Ma NRM is absent in these and more distant unbaked Tudor Volcanics samples.

Point 1 is the strongest evidence and point 4 is the weakest. Together with $^{40}Ar/^{39}Ar$ geochronological evidence that hornblendes in the Tudor Gabbro cooled through ≈590°C around 1100 Ma, we believe that the Tudor A NRM represents a fraction of primary single-domain TRM acquired at ≈1100 Ma that has survived later ≈500°C Grenvillian burial reheating. If this is correct, the Tudor Gabbro A NRM is an ideal candidate for paleointensity determination.

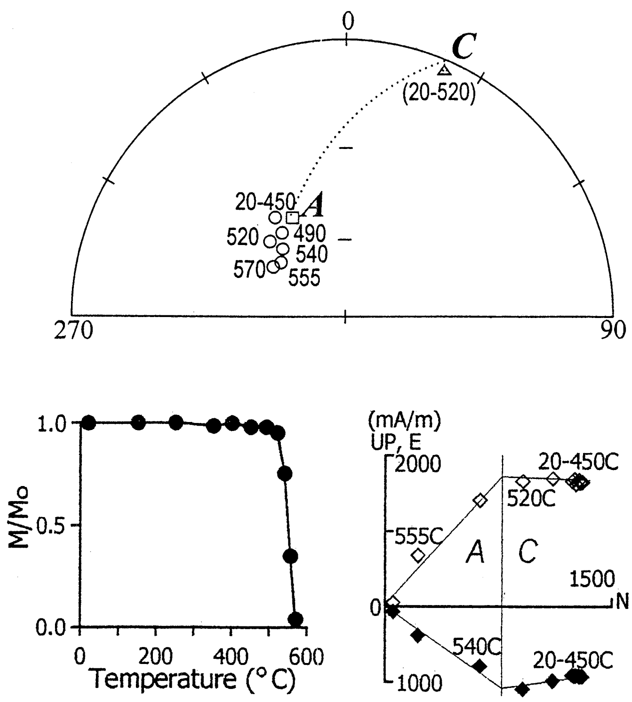

Figure 3. Stereographic, total intensity, and orthogonal vector plots of thermal demagnetization data. Component C is erased from 20 to 520°C (open triangle; (20-520°C) subtracted vector) and component A from 530 to 580°C. The total intensity curve is flat below 520°C because C is approximately perpendicular to A and their vector sum remains almost constant during demagnetization. The junctions between C and A in vertical plane (open diamonds) and horizontal plane (solid triangles) projections are sharp, showing thermoviscous overprinting.

4. Paleointensity Experimental Procedures

The 246 specimens of 45 independently oriented samples from sites 1-9 were divided into three sets for paleointensity experiments. The first set (89 specimens) was heated in air, using the Coe-modified Thellier method [*Coe*, 1967]. After the first heating-cooling step to temperature T_i in field H = 0, the remanence was measured, and the NRM lost in this step was calculated. After the second heating-cooling step to the same temperature T_i with a laboratory field H = 60.0 ± 1.2 μT applied along the specimen axis, the pTRM acquired in cooling from T_i to room temperature T_0 was calculated by taking the vector difference between the remanences measured after the first and second steps. Double heatings were carried out in 50° steps between 300 and 500°C and then in 10°C steps from 500 to 580°C, using an MMTD furnace. Five pTRM checks were made at 450, 500, 520, 540, and 560°C following the 500, 520, 540, 560, and 580°C steps, respectively.

The same procedure was followed for the second set (137 specimens) except that the field applied was 20.0 ± 0.2 μT, chosen for a closer match to the paleointensity values from the first set. Seventy specimens were heated in air, while 67 were heated in helium in an attempt to reduce possible oxidation.

For the third set (20 specimens), "pTRM tail checks" [*Riisager et al.*, 2000] were performed. Following the first (zero-field) and second (in-field) heatings at T_i, samples were heated a third time to T_i in H = 0 to check whether or not the pTRM acquired in the second heating is completely demagnetized by zero-field reheating to its maximum blocking temperature T_i. The difference between the remanences measured after the third and first steps is the pTRM tail, which is expected to be zero for single-domain grains but can be substantial for multidomain grains [*Shcherbakova et al.*, 2000]. Because it remains undemagnetized, the pTRM tail effectively is added to, and contaminates, the NRM for the T_{i+1} and subsequent heating steps [*Dunlop and Özdemir*, 2001]. If the laboratory field **H**, and thus the pTRM tail, are at a large angle to the initial NRM, such contamination may be recognizable by a progressive displacement of the "NRM" direction toward the pTRM direction in higher temperature heatings. Following the pTRM tail check, a fourth in-field heating to a lower temperature T_{i-2} was carried out to obtain a conventional pTRM check.

Throughout all heatings, temperatures were reproducible within ± 2°C. The residual field in the furnace during nominally zero-field heatings and coolings was <150 nT.

To detect possible chemical alteration, bulk magnetic susceptibility was remeasured after heating to each temperature T_i. A modified *Hill and Shaw* [1999] alteration test was also performed. Three or more chips from unused sister specimens were heated along with the paleointensity specimens, and hysteresis loops were measured with an alternating gradient force magnetometer initially and after ≈30-min heatings to 300, 400, 500, 520, 540, 560, and 580°C.

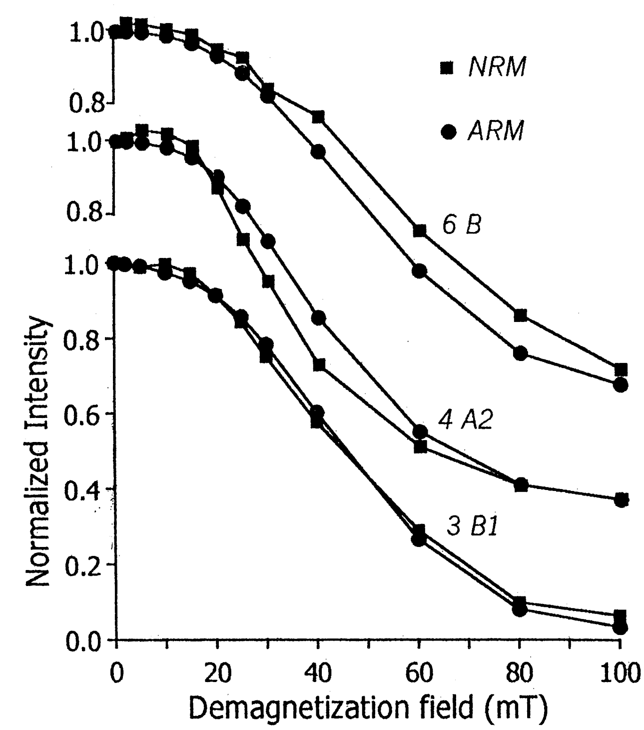

Figure 4. AF demagnetization of NRM and anhysteretic remanence (ARM) in specimens with a dominant A component. The initial plateaus, concentration of coercivities between 30 and 60-80 mT, and doubly curved form of the curves are diagnostic of elongated single-domain or small PSD grains of magnetite.

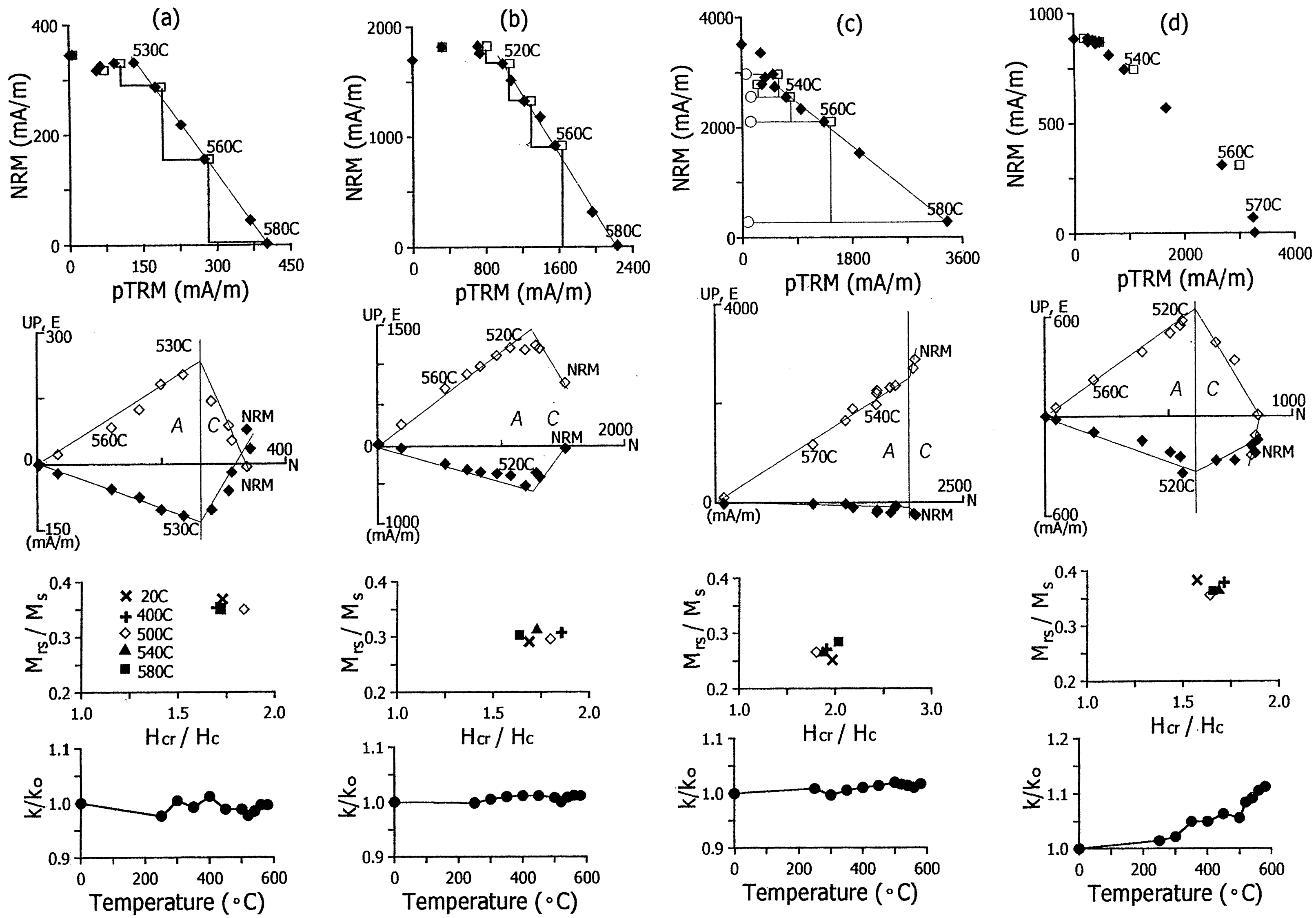

Figure 5. Examples of accepted (a, b, c) 38 3-2, 43 B2, 24 C2 and rejected (d) 40 J3 paleointensity results. For accepted specimens, Arai plots of NRM remaining versus pTRM gained are linear over the range 520 or 530°C to 580°C with pTRM checks (open squares) reproducing the original pTRMs to within 15%, orthogonal vector plots (symbols as in Figure 3) show a clean separation of C and A vectors in thermal demagnetization (zero-field steps in the Thellier experiment), and susceptibility k and hysteresis parameters, such as M_{rs}/M_s and H_{cr}/H_c, remain constant through all heatings. In Figure 5c, open circles are pTRM tail checks carried out as a third zero-field heating step before the conventional pTRM check (fourth step; open squares).

5. Anisotropy Corrections

Magnetic fabric anisotropy can seriously affect paleointensity as well as paleodirectional results [*Rogers et al.*, 1979]. *Selkin et al.* [2000] compared anisotropy of magnetic susceptibility, anisotropy of TRM (ATRM), and anisotropy of ARM (AARM) as indicators of fabric in the ≈2700 Ma Stillwater Complex and devised a method of correcting paleointensities using AARM and/or ATRM.

We used AARM in correcting our paleointensities because determining ATRM requires at least six heatings to 580°C, which may alter magnetic minerals. We produced ARM in six successive directions (±z, ±x, ±y) to obtain the AARM tensor, R_{ij} (i, j = x, y, z). The ARM was imparted in a peak alternating field of 100 mT with an added steady field of 20 μT, using a Schonstedt AF demagnetizer. The corrected remanence vector is $M_i = R_{ij}^{-1} M_j$, where M_j is the uncorrected remanence vector and R_{ij}^{-1} is the inverse of the AARM tensor.

6. Paleointensity Experiments and Results

Paleointensity results were accepted only if they satisfied the following selection criteria.

1. The straight-line portion of the Arai plot [*Nagata et al.*, 1963] must be defined by at least five points, with at least four successive pTRM checks in agreement within 10%.
2. The quality factor S′ obtained from the regression fit to the best straight line must pass statistical rejection criteria [*Yu et al.*, 2000].
3. The NRM must be univectorial over the temperature range used for paleointensity determination, with no tendency to shift its direction (in zero-field thermal demagnetization steps) toward the direction of the field **H** applied during in-field steps. This is a minimum safeguard against contamination by undemagnetized pTRM tails. Where carried out, pTRM tail checks must be null.
4. Bulk susceptibility must remain constant to within ±10% throughout the heating range used for paleointensity determination. Hysteresis parameters should also remain unchanged (however, these proved less sensitive to chemical alteration: see below).

Figures 5a-5c show representative results for accepted specimens, and Figure 5d shows a typical rejected result. Specimen 38 3-2 (Figure 5a) displays a perfectly linear Arai plot over six successive temperature steps, from 530 to 580°C. This is precisely the range over which the vector plot shows univectorial decay of the A component. The pTRM checks reproduce the original 560, 540, 520, and 500°C pTRMs. Both susceptibility and hysteresis parameters, such as M_{rs}/M_s and H_{cr}/H_c, displayed on a Day plot [*Day et al.*, 1977], remain constant within measurement error for all heating steps up to 580°C.

Notice that we could determine a paleointensity value for the C component as well. The decay of the C NRM is not well displayed on a standard Arai diagram because total NRM intensity is plotted as the ordinate. Since the C vector is approximately perpendicular to the A vector for most specimens, one gains the erroneous impression from the Arai diagram that there is no loss of NRM over the 20-520°C range, whereas it is clear that pTRM is being acquired over this range. The key is that pTRM is acquired in a constant direction, the field direction **H**, over the entire temperature span 20-580°C and so is accurately displayed on the abscissa, whereas the NRM has entirely different directions from 20 to 525°C and from 525 to 580°C and is only accurately displayed over the latter interval.

The solution is to measure the incremental decay of the C NRM directly on the vector plot, as proposed by *Dunlop* [1979], and to use these data to construct a separate Arai diagram for the C remanence. This does not seem to have been done in practice for multivectorial NRMs in the literature. We have not carried through a determination of the C paleointensity for our specimens because C was acquired in extremely slow cooling during erosional uplift following burial and the paleointensity determined would be an average over a very long time interval.

Specimen 43 B2 (Figure 5b) has paleointensity data of similar quality to those of 38 3-2. Results for seven successive temperature steps from 520 to 580°C fall on an excellent straight line. Four successive pTRM checks from temperatures as high as 580°C replicate lower-T pTRMs closely. Susceptibility is constant throughout all heatings, as are hysteresis parameters.

Specimen 24 C2 (Figure 5c) displays similar linearity of the Arai plot from 530 to 580°C. In this case, a pTRM tail check was carried out before each pTRM check. There are small positive pTRM tails. The tail at 540°C is marginally significant (≈15% of the pTRM), but it is reduced by heating to the next step, 550°C, and the NRM direction does not change. The pTRM tail of 130 mA/m at 560°C is similar to the difference of 115 mA/m between the 560°C pTRM check and the original pTRM. Thus pTRM tails, not sample alteration, may be the source of pTRM checks that overshoot original pTRMs in our samples. Supporting evidence is the fact that susceptibility and hysteresis parameters are unchanged by heating in most samples, including this one.

The results for specimen 40 J3 (Figure 5d) were rejected. Although the Arai plot is linear from 520-560°C, the 570 and 580°C steps deviate from the linear trend. Chemical alteration is indicated by a steady increase in bulk susceptibility which accelerates above 500°C and by increasing discrepancies in pTRM checks after the 560°C and 580°C heating steps. The hysteresis parameters remain invariant and are not a sensitive indicator of alteration of the NRM carrying fraction.

The 45 accepted paleointensity results before and after anisotropy correction with their associated statistical parameters [*Coe et al.*, 1978; *Yu et al.*, 2000] appear in Table 2. Each paleointensity was determined by leastsquares fitting to the linear portion of the Arai plot, usually between 530°C and 580°C.

The mean paleointensity, without weighting and before AARM correction, is 22.6 ± 3.8 μT. AARM correction changed the mean only slightly, to 23.2 ± 4.3 μT. Weighted mean paleointensity values, following the weighting procedure outlined by *Yu et al.*, [2000], were very similar: 21.9 μT with or without AARM correction. All these values are indistinguishable within their error limits. The distributions of unweighted paleointensity values before and after AARM correction (Figure 6) resemble Gaussian distributions about the respective means H_b and H_a. The effect of anisotropy was not significant: 35 of the 45 accepted values change <10% as a result of AARM correction.

The Tudor gabbro is 3 km × 8 km in areal extent (Figure 2) and must taken a few thousand years to cool through the A NRM blocking temperature range of 580°C to 480-500°C, whereas pTRM was acquired in the laboratory in 0.5-1 hour heatings. The intensities of NRM and pTRM should be adjusted for the differing timescales. Experimental results on the effect of cooling rate on total TRM intensity are

Table 2. Paleointensity Results [a]

Specimen	Site	*Atmosphere*	*N*	ΔT, °C	$H_b \pm \sigma H_b$, μT	$H_a \pm \sigma H_a$, μT	*f*	*g*	*q*	*S'*
1 A1	1	He	6	530-580	24.72 ± 1.24	24.21 ± 1.21	0.59	0.78	9.29	0.287
1 A2	1	Air	7	520-580	18.76 ± 0.66	19.26 ± 0.68	0.71	0.81	16.42	1.961
1 G1	1	Air	5	540-580	25.05 ± 1.97	26.70 ± 2.10	0.37	0.68	3.19	0.137
2 B1	1	He	7	520-580	20.73 ± 0.79	20.56 ± 0.79	0.68	0.81	14.40	0.074
2 B2	1	Air	7	520-580	21.17 ± 0.85	21.38 ± 0.86	0.64	0.77	12.23	0.081
2 C1	1	Air	6	530-580	25.36 ± 1.07	25.45 ± 1.07	0.63	0.76	11.30	0.047
17 B1	2	Air	6	530-580	21.36 ± 0.89	23.69 ± 1.00	0.77	0.75	13.77	0.742
17 B2	2	He	6	530-580	20.73 ± 0.81	20.51 ± 0.81	0.73	0.71	13.30	0.218
17 C1	2	Air	6	530-580	19.92 ± 0.75	20.61 ± 0.77	0.77	0.76	15.61	0.099
19 B1	2	Air	6	530-580	20.58 ± 0.72	20.91 ± 0.74	0.88	0.44	11.05	1.194
22 B2	3	Air	6	530-580	13.92 ± 0.52	14.96 ± 0.35	0.89	0.71	27.20	0.405
24 C2	3	Air	6	530-580	17.77 ± 0.70	18.34 ± 0.72	0.80	0.64	13.28	0.520
30 C1	4	Air	6	530-580	25.61 ± 1.13	25.11 ± 1.11	0.74	0.69	11.60	0.530
32 A1	5	He	5	540-580	19.79 ± 1.35	18.99 ± 1.29	0.43	0.69	4.35	0.197
32 B1	5	Air	5	540-580	21.38 ± 1.14	22.74 ± 1.22	0.57	0.72	7.69	0.090
33 B1	5	Air	6	530-580	21.01 ± 0.92	19.08 ± 0.83	0.65	0.73	10.90	0.624
33 D1	5	He	6	530-580	16.21 ± 0.60	17.37 ± 0.64	0.80	0.74	15.88	0.477
34 B1	5	Air	7	520-580	17.19 ± 0.61	16.04 ± 0.57	0.63	0.77	16.97	0.380
34 C1	5	Air	7	520-580	19.25 ± 0.85	20.16 ± 0.89	0.72	0.72	12.62	0.525
35 A1	5	Air	6	530-580	20.95 ± 1.22	22.33 ± 1.30	0.49	0.77	6.53	0.333
36 B1	6	Air	7	520-580	17.27 ± 0.77	18.00 ± 0.80	0.64	0.82	11.91	0.448
36 E1	6	Air	5	540-580	18.09 ± 0.97	18.13 ± 0.97	0.58	0.74	8.07	0.067
37 1-3	6	Air	5	540-580	15.02 ± 0.65	15.85 ± 0.69	0.46	0.68	5.26	0.189
37 2-1	6	Air	5	540-580	25.94 ± 1.55	26.58 ± 1.59	0.60	0.68	8.62	0.624
37 2-2	6	Air	6	530-580	26.03 ± 1.23	29.41 ± 1.39	0.68	0.71	12.37	0.848
37 3-2	6	He	5	540-580	22.64 ± 1.07	24.16 ± 1.14	0.64	0.71	9.68	0.446
37 3-3	6	Air	6	530-580	27.30 ± 1.40	26.51 ± 1.36	0.53	0.74	7.65	0.205
38 2-2	6	Air	6	530-580	23.60 ± 1.03	24.20 ± 1.06	0.61	0.72	10.09	0.375
38 3-2	6	He	6	530-580	24.41 ± 1.06	24.83 ± 1.06	0.66	0.77	12.68	0.125
38 A1	6	Air	6	530-580	15.02 ± 0.65	15.85 ± 0.69	0.67	0.74	11.45	1.907
43 A1	7	Air	6	530-580	21.36 ± 1.01	21.98 ± 1.04	0.63	0.77	10.25	0.164
43 A2	7	Air	6	530-580	23.94 ± 1.10	27.69 ± 1.29	0.67	0.79	11.50	0.322
43 B1	7	Air	6	530-580	25.63 ± 1.21	27.10 ± 1.29	0.61	0.75	9.75	0.420
43 B2	7	He	7	520-580	24.01 ± 1.20	26.76 ± 1.34	0.55	0.78	8.62	0.337
45 E1	7	Air	6	530-580	22.07 ± 0.99	23.54 ± 1.06	0.71	0.80	12.57	1.113
46 D1	7	He	7	520-580	26.43 ± 1.23	25.30 ± 1.35	0.58	0.83	10.35	0.253
46 E1	7	Air	6	530-580	26.80 ± 1.31	27.31 ± 1.33	0.57	0.79	9.33	0.415
46 E2	7	Air	6	530-580	28.37 ± 1.39	31.63 ± 1.55	0.58	0.78	9.27	0.202
54 1-5	8	Air	6	530-580	23.32 ± 1.07	29.61 ± 2.00	0.66	0.79	11.50	0.303
54 3-2	8	He	6	530-580	29.61 ± 2.00	25.90 ± 1.76	0.43	0.78	5.17	0.468
54 5	8	Air	6	530-580	29.02 ± 1.33	19.70 ± 0.90	0.63	0.78	10.92	0.809
64 A1	9	Air	7	520-580	29.57 ± 2.03	35.55 ± 2.44	0.41	0.81	4.82	0.019
64 B1	9	Air	6	530-580	24.39 ± 2.38	28.42 ± 2.79	0.32	0.79	2.58	0.042
64 B2	9	Air	6	530-580	22.81 ± 1.27	23.61 ± 1.30	0.50	0.76	6.86	0.145
64 C1	9	He	6	530-580	21.32 ± 1.74	24.89 ± 2.03	0.34	0.75	3.19	0.111
					22.58 ± 3.80 [b]	23.15 ± 4.26 [b]				
					21.94 ± 3.85 [c]	21.93 ± 4.44 [c]				

[a] *N* is the number of points used in paleointensity determination; ΔT is the interval of temperature used in slope calculation; $H_b \pm \sigma H_b$ and $H_a \pm \sigma H_a$ are the estimated paleointensity and its standard error from least squares fitting before and after AARM correction; *f*, *g*, and *q* are NRM fraction, gap factor, and quality factor [*Coe et al.*, 1978]; *S'* is the quality factor defined by *Yu et al.* [2000].

[b] unweighted mean calculation.

[c] weighted mean calculation used formulas of *Yu et al.* [2000].

[b,c] meanuncertainties in mean values are standard deviation rather than commonly used standard error expression [*Coe et al.*, 1978].

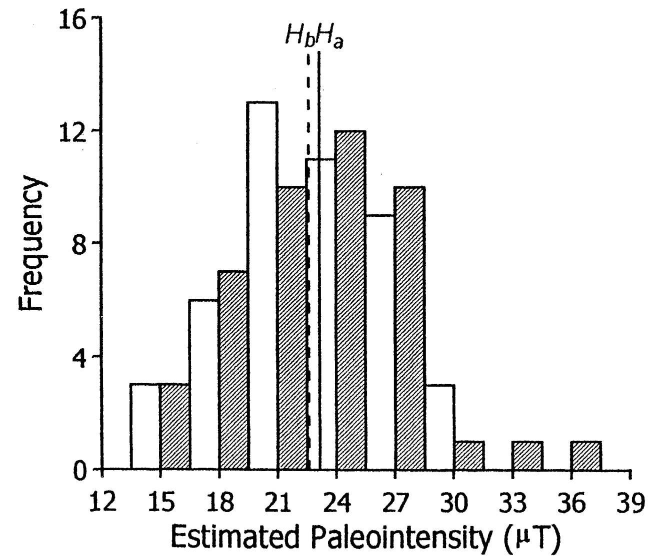

Figure 6. Histograms of paleointensity values before (unshaded bars, mean H_b) and after (shaded bars, mean H_a) AARM correction.

contradictory, however [*Perrin*, 1998], and the effect on partial TRMs in various blocking temperature ranges is unknown. Therefore we did not attempt a coolingrate correction to the paleointensity data.

7. A Paleointensity Simulation Experiment

In order to simulate the paleointensity experiment on the A NRM and test whether the carriers have single-domain character, we carried out the following experiment on the 45 accepted specimens. First, we produced a total TRM by cooling from 600°C in a field of 20 μT. This TRM simulates the original A remanence. Second, we reheated each specimen to 520°C and cooled to room temperature in zero field. This partially demagnetized remanence represents the part of the A that survived reheating in nature to 480-500°C. (We did not cool in a perpendicular field to simulate a C overprint because we are only interested in A.) Finally, we repeated the Thellier paleointensity experiment, as described earlier, using the partially demagnetized remanence as our "NRM" and a field of 20 μT to produce pTRMs.

Arai plots for the Thellier simulation experiment on specimens 38 3-2, 43 B2, and 24 C2 appear in Figures 7a, 7b, and 7c, respectively. The corresponding NRM Arai plots are in Figures 5a, 5b, and 5c. The two sets of Arai plots are practically indistinguishable.

1. There is a sharp junction in the Arai plot at 520°C or 530°C in both cases. An exact match between the maximum unblocking temperature in prior overprinting and the minimum blocking temperature of the surviving remanence is characteristic of single-domain or small pseudo-single-domain (PSD) grains [*Dunlop and Özdemir*, 2000, 2001].
2. The distribution of unblocking temperatures from 530°C to 580°C is identical for NRM and TRM. For example, compare the spacing of points along the Arai line in Figures 5c and 7c (24 C2), particularly the large interval between the 570°C and 580°C points. The spacing of points in Figures 5b and 7b (43 B2) is also distinctive, with a large interval between the 560°C and 570°C points and a small interval between the 570°C and 580°C points. The match between the unblocking temperature distributions is strong evidence that the A NRM is a TRM (and therefore a primary TRM, since the Tudor was never reheated in this temperature range).
3. The 530°C, 540°C, and 560°C pTRM checks for the artificial remanences have exactly the same pattern as the pTRM checks in Figure 5.

We conclude that the A NRM of the Tudor Gabbro is a primary TRM carried by single-domain or small PSD magnetite. Contamination of the NRM by thermochemical remanent magnetization acquired during the paleointensity experiment [*McClelland and Briden*, 1996] seems unlikely both because the minimum blocking/unblocking temperature of the A NRM matches reheating temperatures in nature and because the NRM direction does not move toward the direction of the field applied during pTRM acquisition. Our

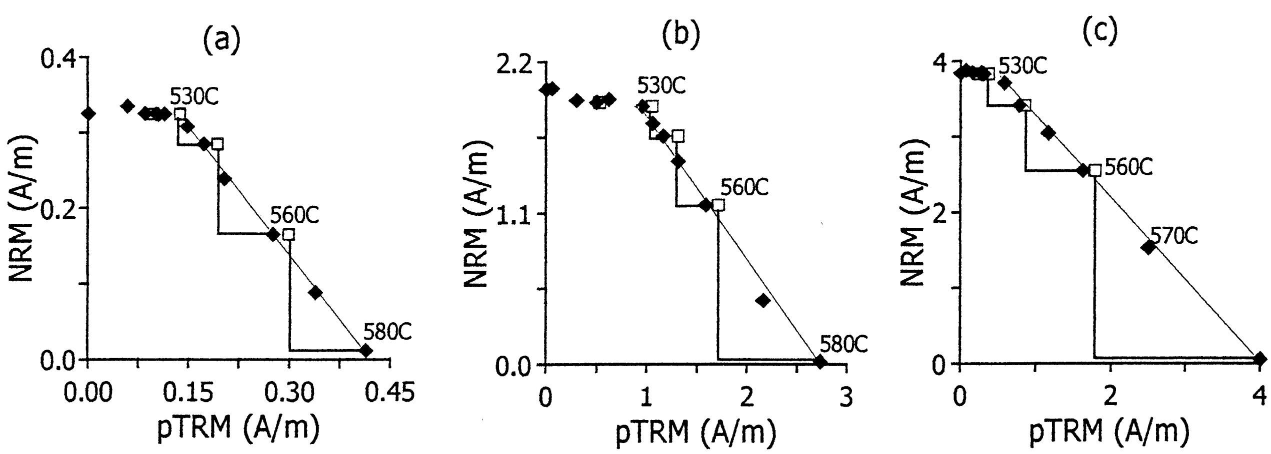

Figure 7. Thellier simulation experiments on specimens (a) 38 3-2, (b) 43 B2, and (c) 24 C2. Instead of NRM, the starting remanence was a laboratory TRM thermally demagnetized to 520°C to imitate the effect of reheating in nature. The Arai plots are very similar to the corresponding NRM paleointensity plots (Figures 5a, 5b, 5c) in their region of linearity (520°C or 530°C to 580°C), the distribution of points along this line during thermal demagnetization, and the pTRM checks, demonstrating that the A NRM is a TRM.

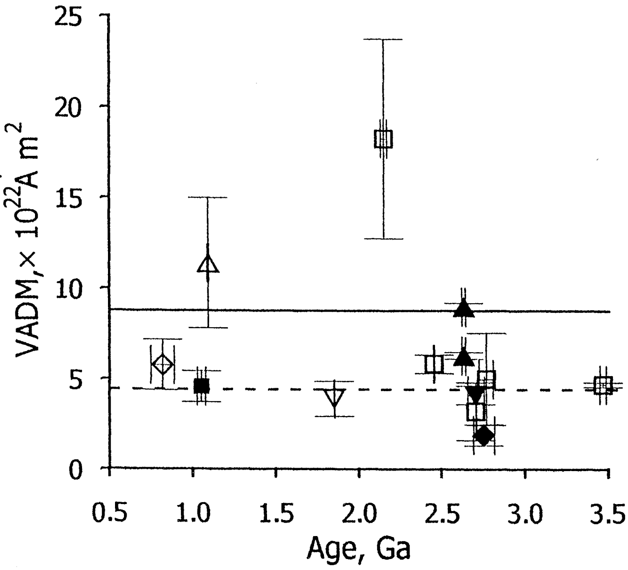

Figure 8. Thellier-type paleointensity determinations for Precambrian rocks. Open squares; *Hale and Dunlop* [1984], *Hale* [1985, 1987]; solid diamond; *Morimoto et al.* [1997]; solid inverted triangle; *Selkin et al.* [2000]; solid triangles; *Yoshihara and Hamano* [2000]; open inverted triangle; *Schwarz and Symons* [1970]; open triangle; *Pesonen and Halls* [1983]; solid circle; *this study*; open diamond; *Schwarz and Symons* [1969]. solid horizontal line; mean dipole moment for the last 10,000 years [*McElhinny and Senanayake*, 1982]; dashed horizontal line; proposed mean Cretaceous-Cenozoic dipole moment [*Juárez et al.*, 1998].

paleointensity estimates therefore record Earth's magnetic field over a period of a few thousand years around 1100 Ma.

8. Earth's Dipole Moment in Precambrian Time

The average inclination I of the A NRM of our 45 accepted specimens is –48°, in good agreement with I = – 46° as found by *Dunlop et al.* [1985]. For a geocentric axial dipole field the corresponding paleolatitude is $\lambda = -29°$. Using this λ and the unweighted mean paleointensity value 23.15 ± 4.26 μT (Table 2) in the dipole field formula $B = \mu_0 H = \mu_0 m (3 \sin^2 \lambda + 1)^{1/2} a^{-3}$, where m is dipole moment and a is the Earth's radius, we find a value m = 4.58 ± 0.84 × 10^{22} A m^2 for Earth's virtual axial dipole moment (VADM). This is about one-half of the mean VADM of 8.8 × 10^{22} A m^2 for the last 10,000 years [*McElhinny and Senanayake*, 1982].

VADMs from all Thellier-type paleointensity determinations for the Precambrian (Table 1) are plotted in Figure 8. Except for the 2150 Ma Preissac dikes and ≈1100 Ma Keweenawan rocks, all are close to or below (in some cases much below) the current VADM. The high VADM of 18.3 × 10^{22} A m^2 for the Preissac dikes was cited by *Hale* [1987] as evidence for onset of a vigorous convective dynamo, perhaps related to nucleation of the inner core, at the close of the Archean (≈2500 Ma). However, the Preissac mean is based on only three specimens and has a large dispersion, whereas the much lower VADM of 3.90 × 10^{22} A m^2 for the only slightly younger (≈1850 Ma) Sudbury Norite is based on 52 specimens and has low dispersion.

There is a similar, although considerably smaller, contrast between our Tudor Gabbro VADM of 4.58 ± 0.84 × 10^{22} A m^2 and the VADM of 11.4 ± 3.6 × 10^{22} A m^2 for Keweenawan rocks [*Pesonen and Halls*, 1983]. Both are nominally the same age, ≈1100 Ma, and both average large numbers of specimens (45 and 54, respectively). The dispersion about the mean is larger for the Keweenawan (≈31%, compared to ≈18% for the Tudor), and no pTRM checks were reported. However, the Keweenawan study was extremely thorough in all other respects (including reversal and baked contact tests), and the Keweenawan pole is one of the key poles for the Precambrian. There seems little doubt that the Keweenawan paleointensity value is trustworthy.

How similar are the Tudor and Keweenawan VADM ages? U/Pb zircon ages indicate a time span of ≈22 Myr for most Keweenawan igneous activity, with the younger ages being about 1085 Ma [*Palmer and Davis*, 1987]. Most of the rocks are volcanic flows or thin intrusions, which must have cooled rapidly and therefore record spot readings of the field. The lack of full averaging of secular variation may be responsible for the larger dispersion about the mean. For the Tudor Gabbro, *Baksi* [1982] reported 1110 ± 20 Ma (1σ error) as the time of cooling through the hornblende Ar blocking temperature of 590°C. This is a hornblende plateau age for a single sample, and although plagioclase and wholerock data from the same heating steps used to define the plateau are in agreement, their error bounds are much larger. The Tudor body also took a substantial length of time to cool through the blocking range from 580°C to 480-500°C: a few thousand years if it was intruded into shallow, cool crust and conceivably, as long as tens of millions of years if it was intruded during the Grenvillian orogeny into surroundings already at ≈500°C [*Baksi*, 1982]. The latter scenario is unlikely, however, because of the large and abrupt change of NRM direction from the older A to the younger C NRM, which records the field during cooling from ≈500°C. A lower limit on the age of the Tudor A is the time of docking of the Elsevir terrane of Grenvillia with Interior Laurentia, marked by the junction in their polar wander paths around 1050 Ma.

On this evidence, both the Keweenawan VADM (11.4 × 10^{22} A m^2) and the Tudor VADM (4.58 × 10^{22} A m^2) appear to be reliable. Each records Earth's field around 1100 Ma with an uncertainty of ~20 Myr. The difference between the two VADM values is at the limit of cyclic variations in dipole field strength in modern times (which should in any case be averaged out over the duration of volcanism or post-intrusive cooling) but is not without precedent in the Phanerozoic. For example, the Rajmahal Traps (113-116 Ma) record a VADM of 12.5 × 10^{22} A m^2 [*Tarduno et al.*, 2001], 3 times the Cenozoic and Late Jurassic-Early Cretaceous average [*Juárez et al.*, 1998].

9. Discussion and Conclusions

Among late Precambrian formations of the Grenville Province, the Tudor Gabbro is a particularly favorable candidate for paleointensity determination. Its metamorphic grade is the lowest found anywhere in the Grenville. Garnet-pyroxene geothermometry [*Anovitz and Essene*, 1990] indicates that it was never reheated above 480-500°C. Its NRM consists of two components, an A NRM with laboratory unblocking temperatures between 530°C and 580°C and a C NRM of very different direction with unblocking temperatures entirely below 520-530°C. The junction between the two vectors is sharp in thermal demagnetization, suggesting that C is a thermoviscous overprint of A. Indeed, after allowing for the difference in heating/cooling timescales in nature and in the laboratory [*Pullaiah et al.*, 1975], the junction temperature corresponds exactly to the peak metamorphic reheating temperature of 480-500°C. Therefore, following *Dunlop et al.* [1985], we interpret A to be the surviving part of a premetamorphic primary TRM and C to be the record of thermoviscous remagnetization during very slow cooling caused by erosional unroofing and uplift of the Grenville orogen.

The primary TRM nature of the Tudor Gabbro A is inferred from several lines of evidence. A cannot be a late CRM because C clearly overprints A. It cannot be a CRM contemporaneous with 950-800 Ma uplift and cooling because the A direction does not fit the established Grenville Track. It is unlikely to be an early (pre-950 Ma) CRM because the A direction is most consistent with the age interval 1100-1050 Ma, predating Grenvillian collision but matching the intrusion age of ≈1100 Ma [*Baksi*, 1982]. A also cannot be, even in part, a laboratory-acquired CRM or TCRM because of the sharp segregation of its blocking temperatures above 520-530°C and because the NRM direction remains unchanged, although fields are applied at large angles to A to produce pTRM.

Furthermore, the AF demagnetization curve of A has the distinctive sigmoid shape and high coercivities associated with elongated single-domain or small PSD grains of magnetite. The carrier of the A is inferred to be submicroscopic magnetite needles or rods exsolved in plagioclase and/or pyroxene crystals during primary cooling at temperatures above the magnetite Curie point. The single-domain character of the A NRM makes it ideal for paleointensity work. A direct comparison of the unblocking temperature distributions of the A NRM and of laboratory TRM (Figures 5 and 7) shows a good match. On this magnetic and mineralogical evidence, A is a primary TRM.

Reliable paleointensity results that pass stringent selection criteria were obtained for 45 specimens from 19 samples at 9 sites. Anisotropy was not significant in most cases but was, nevertheless, corrected for using the AARM tensor determined for each specimen. The final mean paleointensity was 23.2 ± 4.3 μT. From the paleolatitude of −29° indicated by the mean A inclination of −48°, a VADM of 4.58 ± 0.84 μT was calculated for the Earth's dipole field around 1100 Ma. Since the Tudor Gabbro must have taken at least a few thousand years to cool from 580°C to 480-500°C following its intrusion, the paleomagnetic data average out secular variation, including cyclic variations in paleofield strength.

A much higher VADM of 11.4 ± 3.6 μT for Keweenawan rocks of about the same age [*Pesonen and Halls*, 1983] also seems to be reliable. The difference in time between the two is probably no more than 20 Myr. Similar contrasts in paleointensity are documented for similar time intervals in the Phanerozoic [*Tarduno et al.*, 2001] and seem to be a feature of the late Archean and Proterozoic (Figure 8). However, the hypothesis of a sudden onset of vigorous convection at the close of the Archean with an accompanying jump in geomagnetic field strength [*Hale*, 1987] needs to be reexamined because it is based on a single high paleointensity value for the 2150 Ma Preissac dikes, averaging only three specimens with large dispersion. Most later paleointensities, including the ≈1850 Ma Sudbury Norite (52 specimens), are much lower and resemble both earlier (Archean) and later (Cretaceous-Cenozoic) values. If there was a dynamo spurt in the early Proterozoic, it must have been transitory.

Acknowledgments. We thank Özden Özdemir for useful discussions and Avto Goguitchaichvili, Carlo Laj, and an anonymous referee for helpful reviews that greatly improved the paper. This research has been supported by the Natural Science and Engineering Research Council of Canada through grant A7709 to D.J.D.

References

Anovitz, L. M. , and E. J. Essene, Thermobarometry and pressure-temperature paths in the Grenville Province of Ontario, *J. Petrol.*, *31*, 197-241, 1990.

Baksi, A. K., $^{40}Ar/^{39}Ar$ incremental heating studies on the Tudor Gabbro, Grenville Province, Ontario: Its bearing on the North America apparent polar wander path in late Proterozoic times, *Geophys. J. R. Astronon. Soc.*, *70*, 545-562, 1982.

Berger, G. W., and D. York, $^{40}Ar/^{39}Ar$ dating of the Thanet Gabbro, Ontario: Looking through the Grenvillian metamorphic veil and implications for paleomagnetism, *Can. J. Earth Sci.*, *18*, 266-273, 1981.

Berger, G. W., D. York, and D. J. Dunlop, Calibration of Grenvillian palaeopoles by $^{40}Ar/^{39}Ar$ dating, *Nature*, *277*, 46-48, 1979.

Buchan, K. L., Magnetic overprinting in the Thanet gabbro complex, *Can. J. Earth Sci.*, *15*, 1407-1421, 1978.

Coe, R. S., Paleointensities of the Earth's magnetic field determined from Tertiary and Quaternary rocks, *J. Geophys. Res.*, *72*, 3247-3262, 1967.

Coe, R. S., C. S. Grommé, and E. A. Mankinen, Geomagnetic paleointensities from radiocarbon-dated lava flows on Hawaii and the question of the Pacific nondipole low, *J. Geophys. Res.*, *83*, 1740-1756, 1978.

Cosca, M. A., J. F. Sutter, and E. J. Essene, Cooling and inferred uplift/erosion history of the Grenville Orogen, Ontario: Constraints from $^{40}Ar/^{39}Ar$ thermochronology, *Tectonics*, *10*, 959-977, 1991.

Cosca, M. A., E. J. Essene, M. J. Kunk, and J. F. Sutter, Differential unroofing within the Central Metasedimentary Belt of the Grenville Orogen: Constraints from $^{40}Ar/^{39}Ar$ thermochronology, *Contrib. Mineral. Petrol.*, *110*, 211-225, 1992.

Costanzo-Alvarez, V., and D. J. Dunlop, A regional paleomagnetic study of lithotectonic domains in the Central Gneiss Belt, Grenville Province, Ontario, *Earth Planet. Sci. Lett.*, *157*, 89-103, 1998.

Culshaw, N. G., R. A. Jamieson, N. Wodicka, and J. W. F. Ketchum, Was the Central Metasedimentary Belt emplaced over the Central Gneiss Belt at ~1190 Ma? (abstract), *Geol. Assoc. of Can.*, Waterloo, Ont., 1994.

Dallmeyer, R. D., and J. F. Sutter, Acquisitional chronology of remanent magnetization along the "Grenville polar path": Evidence from $^{40}Ar/^{39}Ar$ ages of hornblende and biotite from the Whitestone diorite, Ontario, *J. Geophys. Res.*, *85*, 3177-3186, 1980.

Davis, K. E., Magnetite rods in plagioclase as the primary carrier of stable NRM in ocean floor gabbros, *Earth Planet. Sci. Lett.*, *55*, 190-198, 1981.

Day, R., M. Fuller, and V. A. Schmidt, Hysteresis properties of titanomagnetites: Grain-size and compositional dependence, *Phys. Earth Planet. Inter.*, *13*, 260-267, 1977.

Dudas, F. O., J. B. Henderson, and J. K. Mortensen, U-Pb ages of zircons from the Anton Complex, southern Slave Province, NWT: Radiogenic ages and isotope studies, *Bull. Geol. Surv. Can.*, *90-2*, 39-44, 1990.

Dunlop, D. J., On the use of Zijderveld vector diagrams in multicomponent paleomagnetic studies, *Phys. Earth Planet. Inter.*, *20*, 12-24, 1979.

Dunlop, D. J., and Ö. Özdemir, *Rock Magnetism: Fundamentals and Frontiers*, 573 pp., Cambridge Univ. Press, New York, 1997.

Dunlop, D. J., and Ö. Özdemir, Effect of grain size and domain state on thermal demagnetization tails, *Geophys. Res. Lett.*, *27*, 1311-1314, 2000.

Dunlop, D. J., and Ö. Özdemir, Beyond Néel's theories: Thermal demagnetization of narrow-band partial thermoremanent magnetizations, *Phys. Earth Planet. Inter.*, in press, 2001.

Dunlop, D. J., and M. Prévot, Magnetic properties and opaque mineralogy of drilled submarine intrusive rocks, *Geophys. J. R. Astron. Soc.*, *69*, 763-802, 1982.

Dunlop, D. J., and J. M. Stirling, Post-tectonic magnetizations from the Cordova gabbro, Ontario and Palaeozoic reactivation in the Grenville Province, *Geophys. J. R. Astron. Soc.*, *82*, 521-550, 1985.

Dunlop, D. J., H. Hyodo, T. Knight, and A. G. Steele, Paleomagnetism of the Tudor gabbro, Ontario: Evidence for divergence between Grenvillia and interior Laurentia, *Geophys. J. R. Astron. Soc.*, *83*, 699-720, 1985.

Dunlop, D. J., P. W. Schmidt, Ö. Özdemir, and D. A. Clark, Paleomagnetism and paleothermometry of the Sydney Basin 1. Thermoviscous and chemical overprinting of the Milton Monzonite, *J. Geophys. Res.*, *102*, 27,271-27,283, 1997a.

Dunlop, D. J., Ö. Özdemir, and P. W. Schmidt, Paleomagnetism and paleothermometry of the Sydney Basin 2.Origin of anomalously high unblocking temperatures, *J. Geophys. Res.*, *102*, 27,285-27,295, 1997b.

Dunlop, D. J., Ö. Özdemir, D. A. Clark, and P. W. Schmidt, Time-temperature relations for the remagnetization of pyrrhotite (Fe_7S_8) and their use in estimating paleotemperatures, *Earth Planet. Sci. Lett.*, *176*, 107-116, 2000.

Easton, R. M., The Grenville Province and the Proterozoic history of central and southern Ontario, *Geology of Ontario*, edited by P. C. Thurston *et al.*, *Ontario Geol. Surv.*, Spec. Vol. 4, 715-904, 1992.

Evans, M. E., and M. L. Wayman, An investigation of small magnetic particles by means of electron microscopy, *Earth Planet. Sci. Lett.*, *9*, 365-370, 1970.

Evans, M. E., and M. L. Wayman, An investigation of the role of ultra-fine titanomagnetite intergrowths in palaeomagnetism, *Geophys. J. R. Astron. Soc.*, *36*, 1-10, 1974.

Garrison, J. R., and L. A. Taylor, Petrogenesis of pyroxene-oxide intergrowths from kimberlite and cumulate rocks: Co-precipitation or exsolution?, *Am. Mineral.*, *66*, 723-740, 1981.

Hale, C. J., Evidence of the Archean geomagnetic field, Ph.D. thesis, 188 pp., Univ. of Toronto, Toronto, Ont., Canada, 1985.

Hale, C. J., Paleomagnetic data suggest link between the Archean-Proterozoic boundary and inner-core nucleation, *Nature*, 329, 233-237, 1987.

Hale, C. J., and D. J. Dunlop, Evidence for an early Archean geomagnetic field: A paleomagnetic study of the Komati formation, Barberton belt, South Africa, *Geophys. Res. Lett.*, *11*, 97-100, 1984.

Hanes, J. A., and D. York, A detailed $^{40}Ar/^{39}Ar$ study of an Abitibi dike from the Canadian Superior Province, *Can. J. Earth Sci.*, 16, 1060-1070, 1979.

Hargraves, R. B., and W. M. Young, Source of stable remanent magnetism in Lambertville diabase, *Am. J. Sci.*, *267*, 1161-1177, 1969.

Heaman, L. M., U-Pb dating of mafic dyke swarms: what are the options? (abstract), *Bull. N. M. Bur. Mines Miner. Res.*, *131*, 125, 1995.

Henderson, J. B., O. Breeman, and W. D. Loveridge, Some U-Pb ages from Archean basement, supracrustal and intrusive rocks, Yellowknife-Hearne Lake area, District of Mackenzie, NWT: Radiogenic age and isotope studies, *Bull. Geol. Surv. Can.*, *87-2*, 111-121, 1987.

Hill, M. J., and J. Shaw, Paleointensity results for historic lavas from Mt. Etna using microwave demagnetization/remagnetization in a modified Thellier-type experiment, *Geophys. J. Int.*, *139*, 583-590, 1999.

Juárez, M. T., L. Tauxe, J. S. Gee, and T. Pick, The intensity of the Earth's magnetic field over the past 160 million years, *Nature*, *394*, 878-881, 1998.

Krogh, T. E., D. W. davis, and F. Corfu, Precise U=Pb zircon and baddeleyite ages for the Sudbury area, *The Geology and Ore Deposits of the Sudbury Structure*, edited by E. G. Pye, A. J. Naldrett, and P. E. Giblin, *Ont. Geol. Surv. Spec. Vol.* 1, 431-446, 1984.

Lopez-Martinez, M., and D. York, Further thermochronometric unraveling of the age and paleomagnetic record of the southwest Grenville Province, *Can. J. Earth Sci.*, *20*, 953-960, 1983.

Lumbers, S. B., Geology of Limerick and Tudor Townships, *67*, 110 pp., *Ont. Dept. Mines, Geol. Rep.*, 1969.

McClelland, E., and J. C. Briden, An improved methodology for Thellier-type paleointensity determination in igneous rocks and its usefulness for verifying primary thermoremanence, *J. Geophys. Res.*, *101*, 21,995-22,013, 1996.

McClelland-brown, E., Discrimination of TRM and CRM by blocking-temperature spectrum analysis, *Phys. Earth Planet. Inter.*, *30*, 405-414, 1982.

McElhinny, M. W., and W. E. Senanayake, Variations in the geomagnetic dipole 1.The past 50,000 years, *J. Geomagn. Geoelectr.*, *34*, 39-51, 1982.

Mezger, K., B. A. van der Pluijm, E. J. Essene, and A. N. Halliday, Synorogenic collapse: A perspective from the middle crust, the Proterozoic Grenville Orogeny, *Science*, *254*, 695-698, 1991.

Morimoto, C., Y. Otofuji, M. Miki, H. Tanaka, and T. Itaya, Preliminary paleomagnetic results of an Archean dolerite dyke of west Greenland: Geomagnetic field intensity at 2.8 Ga, *Geophys. J. Int.*, *128*, 585-593, 1997.

Nagata, T., Y. Arai, and K. Momose, Secular variation of the geomagnetic total force during the last 5000 years, *J. Geophys. Res.*, *68*, 5277-5281, 1963.

Nunes, P. D., and L. S. Jensen, Geochronology of the Abitibi metavolcanic belt, Kirkland lake area - Progress report, *Summary of Geochronology Studies 1977-1979*, edited by E. G. Pye, *Ont. Geol. Surv. Misc. Paper*, *92*, 56 pp., 1980.

Palmer, H. C., and D. W. Davis, Paleomagnetism and U-Pb geochronology of volcanic rocks from Michipicoten Island, Lake Superior, Canada: Precise calibration of the Keweenawan polar wander track, *Precambrian Res.*, *37*, 157-171, 1987.

Perrin, M., Paleointensity determination, magnetic domain structure, and selection criteria, *J. Geophys. Res.*, *103*, 30,591-30,600, 1998.

Pesonen, L. J., and H. C. Halls, Geomagnetic field intensity and reversal asymmetry in late Precambrian Keweenawan rocks, *Geophys. J. R. Astron. Soc.*, *73*, 241-270, 1983.

Premo, W., R. Helz, M. Zientek, and R. Langston, U/Pb and Sm/Nd ages for the Stillwater Complex and its associated sills and dikes, Beartooth Mountains, Montana: Identification of a parent magma?, *Geology, 18*, 1065-1068, 1990.

Prévot, M., and M. Perrin, Intensity of the Earth's magnetic field since Precambrian from Thellier-type palaeointensity data and inferences on the thermal history of the core, *Geophys. J. Int.*, *108*, 613-620, 1992.

Pullaiah, G., E. Irving, K. L. Buchan, and D. J. Dunlop, Magnetization changes caused by burial and uplift, *Earth Planet. Sci. Lett.*, *28*, 133-143, 1975.

Riisager, J., M. Perrin, P. Riisager, and G. Ruffet, Paleomagnetism, paleointensity, and geochronology of Miocene basalts and baked sediments from Velay Oriental, French Massif Central, *J. Geophy. Res.*, *105*, 883-896, 2000.

Rivers, T., J. Martignole, C. F. Gower, and A. Davidson, New tectonic divisions of the Grenville Province, southeast Canadian Shield, *Tectonics*, *8*, 63-84, 1989.

Rogers, J., J. M. W. Fox, and M. J. Aitken, Magnetic anisotropy in ancient pottery, *Nature, 277*, 644-646, 1979.

Schwarz, E. J., and D. T. A. Symons, Geomagnetic intensity between 100 million and 2500 million years ago, *Phys. Earth Planet. Inter.*, *2*, 11-18, 1969.

Schwarz, E. J., and D. T. A. Symons, Paleomagnetic field intensity during cooling of the Sudbury irruptive 1700 million years ago, *J. Geophys. Res.*, *75*, 6631-6640, 1970.

Selkin, P. A., and L. Tauxe, Long-term variations in paleointensity, *Philos. Trans. R. Soc. London Ser. A*, 358, 1065-1088, 2000.

Selkin, P. A., J. S. Gee, L. Tauxe, W. P. Meurer, and A. Newell, The effect of remanence anisotropy on paleointensity estimates: A case study from the Archean Stillwater Complex, *Earth Planet. Sci. Lett.*, *183*, 403-416, 2000.

Shcherbakova, V. V., V. A. Shcherbakov, and F. Heider, Properties of partial thermoremanent magnetization in pseudosingle domain and multidomain magnetite grains, *J. Geophys. Res.*, *105*, 767-781, 2000.

Tarduno, J. A., R. D. Cottrell, and A. V. Smirnov, High geomagnetic intensity during the mid-Cretaceous from Thellier analyses of single plagioclase crystals, *Science*, *291*, 1779-1783, 2001.

Wanless, R. K., R. D. Stevens, G. R. Lachance, and J. Y. H. Rimsaite, Age determinations and geological studies: K/Ar isotope ages, *Pap. Geol. Surv. Can.*, *65-17*, 101 pp., 1966.

Weil, A. B., R. Van der Voo, C. MacNiocaill, and J. G. Meert, The Proterozoic supercontinent Rodinia: Paleomagnetically derived reconstructions for 1100 to 800 Ma, *Earth Planet. Sci. Lett.*, *154*, 13-24, 1998.

Yoshihara, A., and Y. Hamano, Intensity of the Earth's magnetic field in late Archean obtained from diabase dikes of the Slave Province, Canada, *Phys. Earth Planet. Inter.*, *117*, 295-307, 2000.

Yu, Y., D. J. Dunlop, L. Pavlish, and M. Cooper, Archeomagnetism of Ontario potsherds from the last 2000 years, *J. Geophys. Res.*, *105*, 19,419-19,433, 2000.

Zhang, B., and H. C. Halls, The origin and age of feldspar clouding in the Matachewan dyke swarm, Canada, *Physics and Chemistry of Dykes*, edited by G. Baer and A. Heimann, pp. 171-176, A. A. Balkema, Brrokfield, Vt., 1995.

Zindler, A., C. Brooks, N. T. Arndt, and S. R. Hart, Nd and Sr isotope data from komatiitic and tholeiitic rocks of Munro Township, Ontario; *U. S. Geol. Surv. Open File Rep. 78-701*, 469-471, 1978.

Y. Yu and D. J. Dunlop, Geophysics, Department of Physics, University of Toronto at Mississauga, 3359 Mississauga Road North, Mississauga, Onatario, Canada L5L 1C6. (yjyu@physics.utoronto.ca; dunlop@physics.utoronto.ca)

(Received September 29, 2000; revised June 30, 2001; accepted June 30, 2001.)