Paleomagnetism and magnetic mineralogy of Grenville metamorphic and igneous rocks, Adirondack Highlands, USA

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Paleomagnetism and magnetic mineralogy of Grenville metamorphic and igneous rocks, Adirondack Highlands, USA

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ABSTRACT

Paleomagnetic samples were collected from an east–west traverse across the Highlands region of the Adirondack Mountains, northern New York, which forms part of the southern most exposure of the Grenville Province of North America. Granulite facies metamorphism ($T > 650\,^{\circ}$C) at ~1050 Ma completely reset pre-existing magnetic directions in sampled microcline gneisses and metamorphosed anorthosites. Fourteen sites of microcline gneiss yield a mean direction of $I = -62.8^{\circ}$, $D = 289.2^{\circ}$ with $\alpha_{95} = 7.6^{\circ}$ and a corresponding pole at 18.4°S and 151.1°E. Metamorphic anorthosites and associated rocks ($N = 14$) show a direction of $I = -67.3^{\circ}$, $D = 283.9^{\circ}$ with $\alpha_{95} = 7.7^{\circ}$ and a corresponding pole at 25.1°S and 149.0°E. Post-metamorphic fayalite granites possess a statistically different direction of $I = -75.8^{\circ}$ and $D = 297.0^{\circ}$ ($\alpha_{95} = 3.9^{\circ}$, $N = 8$ sites) and a pole of 28.4°S and 132.7°E. Both normal and reverse polarities are recorded in all units, with reverse sites occurring on the eastern and western ends of the traverse, and normal polarities restricted to the central part of the massif. The remanence is carried by ilmeno-hematite in the microcline gneisses. In the metamorphosed anorthosites, and the unmetamorphosed fayalite granites, hemo-ilmenite and magnetite occur, though magnetite is the predominant oxide. Using cooling curves established for the Highlands, and blocking temperatures determined for ilmeno-hematite, hemo-ilmenite and magnetite, the age of remanence is determined to be ~990 Ma for the magnetite-bearing Wanakena granite, ~970 Ma for the metamorphosed anorthosite and related rocks, and ~960 Ma for the ilmeno-hematite rich microcline gneiss. The pole data from the Adirondacks, as well as selected studies from other areas of the Grenville Province on units with similar mineralogy and some age control, helps define the southerly part of the Grenville loop of the apparent polar wander path. The three units from the Adirondacks indicate counter-clockwise motion of the APWP between 990 and 960 Ma.

1. Introduction

The development of an apparent polar wander path (APWP) for a continental craton is a key to understanding the evolution of that continent and its relationship to others. Since the first crude paths were compiled in the 1950s we have been using APWPs to help reconstruct the world in the geologic past. Although paleomagnetic pole positions are relatively straightforward to determine from unaltered rocks on stable cratons, poles from orogenic belts provide special challenges, including the multi-component magnetizations commonly found in metamorphosed rocks, and the difficulty in determining the age of magnetization of the rocks with sufficient accuracy.

During the late Proterozoic the landmass of Laurentia formed a central part of the supercontinent Rodinia (Hoffman, 1991). To examine the formation and development of Rodinia the apparent polar wander paths of various cratons have been compared. The APWP for Laurentia in the late Proterozoic (1200–900 Ma) is a good example of both the power and the problems with pole paths. There have been numerous discussions of the path's shape and form, although the position of the poles in the South Pacific-Australia area and the presence of a loop in the path have been established for some time (Dubois, 1962; Roy, 1983). The 'Grenville loop' derived from Laurentia pole data in the age range of 1100–700 Ma is generally assumed to exhibit clockwise motion (e.g. Alvarez and Dunlop, 1998), although the data has also been interpreted as indicating counterclockwise motion (Weil et al., 1998). The determination of a detailed and accurate APWP is necessary to solve problems of location and motion of other cratons with respect to Laurentia, when these cratons joined or left the Rodinia configuration, and how long the supercontinent of Rodinia survived. Despite the importance of establishing accurate APW paths, determining the

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age of remanence in rocks from orogenic belts to the degree needed to construct temporal relationships between different pole positions is extremely difficult. In the designation of "key poles"—ones that have both stable remanence whose primary nature can be established by field tests and accurate age determinations of the remanence (Buchan et al., 2000), there are few acceptable poles from metamorphic rock units.

Here we present detailed paleomagnetic studies on three rock units from the Adirondack Highlands of New York, the southeasterly extension of exposure of the Grenville orogenic belt. Units sampled include sillimanitic microcline gneisses of the far western Highlands, metamorphosed anorthosites of the Marcy Massif and associated bodies, all metamorphosed to granulite grade, and samples from post-orogenic fayalite granites. The times of acquisition of remanence in the different rock units are estimated from cooling curves derived from published geochronology on associated rocks, and correlated magnetic oxide phases and exsolution microstructures.

2. Geologic setting

The Grenville Province of eastern North America, extending from southern Labrador in the northeast to Lake Huron in the southwest, is a region of late Mesoproterozoic tectonic plutonism and regional metamorphism. The region has long been divided into a series of belts with different ages, compositions, and metamorphic grades (Davidson, 1995; Rivers, 1997, 2008) and has given its name to extensive and complex orogenic activity $\sim$1.0 Ga. Multiple cratons were involved in Grenville-age orogenic activity, from Scandinavia to the complex Australian region with 5 cratons, with the end product being the assembly of the supercontinent of Rodinia (Hoffman, 1991). In the Grenville Province itself, this orogenic activity represented several distinct events considered typical of hot orogens of long duration (Rivers, 2008).

The Adirondack Mountains of New York form the southeast extension of the Grenville Province of Ontario and Quebec, and, with the Morin area north of the St. Lawrence Valley, constitute the Central Granulite Belt (Wynne-Edwards, 1972; McLelland et al., 1996). The Adirondacks are divided into the Highland Province, dominated by granulite-facies metamorphosed igneous rocks including the suite of anorthosite and associated mangerite, charnockite and granite rocks (AMCG), and the Lowlands Province (LP). The LP is mainly made of metamorphosed sedimentary and volcanic rocks, which are dominantly of amphibolite facies. A distinct zone of intensely deformed rocks referred to as the Carthage-Colton shear zone (CCSZ) lies near the boundary of these two regions (Mezger et al., 1992) (Fig. 1). Unlike Grenville basement exposed in the Appalachians to the east, the Adirondacks have not suffered discernible Paleozoic metamorphism or deformation, and thus represent mineral assemblages and structural configurations related to the Grenville orogenic cycle only (McLelland et al., 1996).

The extensive AMCG suite of rocks forming the bulk of the Adirondack Highlands were emplaced into orthogneiss country rocks during the Elzevirian Orogeny recently dated using U–Pb isotopes on zircons, at ca 1.15 Ga (Hamilton et al., 2004; McLelland et al., 2004). Subsequent metamorphism and deformation of the AMCG suite and associated rocks in the Highlands occurred some 60–100 myr later, as part of the Ottawan orogeny (McLelland et al., 2004; Heumann et al., 2006; Bickford et al., 2008). At this time metamorphic conditions for the Highland rocks reached granulite grade, with peak metamorphic temperatures up to 800 °C, and pressures of up to 8 kbar (Bohlen et al., 1985; Spear and Markussen, 1997; Bickford et al., 2008), followed by a long period of orogenic collapse and complex cooling.

The samples collected for this study, in an east–west transect across the Highlands, consist of metamorphic gneisses, metamorphosed anorthosites and related rocks, and post-orogenic granites (Fig. 1). One set of samples is from microcline gneisses (GMS), commonly containing sillimanite and locally garnet, and associated metamorphosed sedimentary layers. GMS samples are located southeast of the Carthage-Colton shear zone in the western to central part of the Highland Province. Although some sillimanite-microcline gneisses are rich in magnetite to the extent of being mined for iron ore (Buddington and Leonard, 1962), other parts of the unit are virtually magnetite-free with ilmeno-hematite (titanohematite with exsolution lamellae of ilmenite ± rutile) as the sole, or dominant iron oxide (Balsley and Buddington, 1957, 1958). Rocks from this unit, the "Russell Belt" (location R in Fig. 1) described by Balsley and Buddington (1957), were studied in detail by McEnroe and Brown (2000) for their rock magnetic properties and the strong association of negative natural remanent magnetization (NRM) directions with a large negative aeromagnetic anomaly. Other locations for GMS samples are from the Oswegatchie area south east of Russell (OSW; O in Fig. 1) and from the Sabattis area south and east of Cranberry Lake (S in Fig. 1).

The second group of samples is from the massive metamorphosed anorthosites of the Marcy Massif (Buddington, 1968), so characteristic of the Adirondack terrain. The Marcy Massif is a complex intrusion with coarse anorthosite and leuconorite grading to more mafic compositions near the borders. Rocks in the center of the massif generally consist of coarse blue-gray plagioclase in a matrix with lighter plagioclase, clinopyroxene, orthopyroxene and minor opaques, hornblende and garnet (Buddington, 1939). Facies closer to the massif edges are typically altered by assimilation of country rock and are commonly recrystallized and finer-grained than the core rocks.

The Wanakena granite is a small body cropping out west of Cranberry Lake in the western Highlands (W in Fig. 1), described by Buddington (1939), Buddington and Leonard (1962) and Jaffe et al. (1978). It is dark green, massive and undeformed fayalite–ferrohedenbergite granite that shows no penetrative tectonic fabric and crosscuts older rocks and fabrics. Buddington and Leonard (1962) consider it the youngest igneous rock in the Adirondacks except for later diabase dikes. McLelland et al. (2001) classify the Wanakena as part of the Lyon Mountain Granitic suite and regard it as a terminal, post-orogenic intrusion. Recent zircon geochronology places the Wanakena at 1047 ± 10 Ma, after the peak metamorphism experienced in the Adirondacks (McLelland et al., 2001). Another post-metamorphic fayalite granite, cropping out near Ausable Forks, NY, which is very similar in composition to the Wanakena, was also sampled (A in Fig. 1). This body is mapped as intruding all the other rocks in the region (Whitney and Olmsted, 1993) and has been dated at 1047 ± 2.2 Ma (McLelland et al., 2001), the same age, but with tighter control, as the Wanakena intrusion.

2.1. Sampling

Metamorphic rocks of the Adirondack Highlands were sampled in an east–west traverse, from the Carthage-Colton mylonite zone in the west to Lake Champlain in the east (Fig. 1). The microcline-sillimanite gneiss (GMS unit) was sampled in the Russell and Oswegatchie (OSW) areas in the far-western Highlands as well as in the Sabattis and Sevey areas in the central northwestern Adirondacks (Fig. 1). Samples for the Wanakena granite were all collected west of the town of Wanakena along State Highway 3. Anorthosite sites were located across the Marcy massif from near Saranac Lake eastward to Elizabethtown and Lake Champlain, as well as several sites from outlying bodies south of the main anorthosite unit. In addition, a number of other units were sampled, including

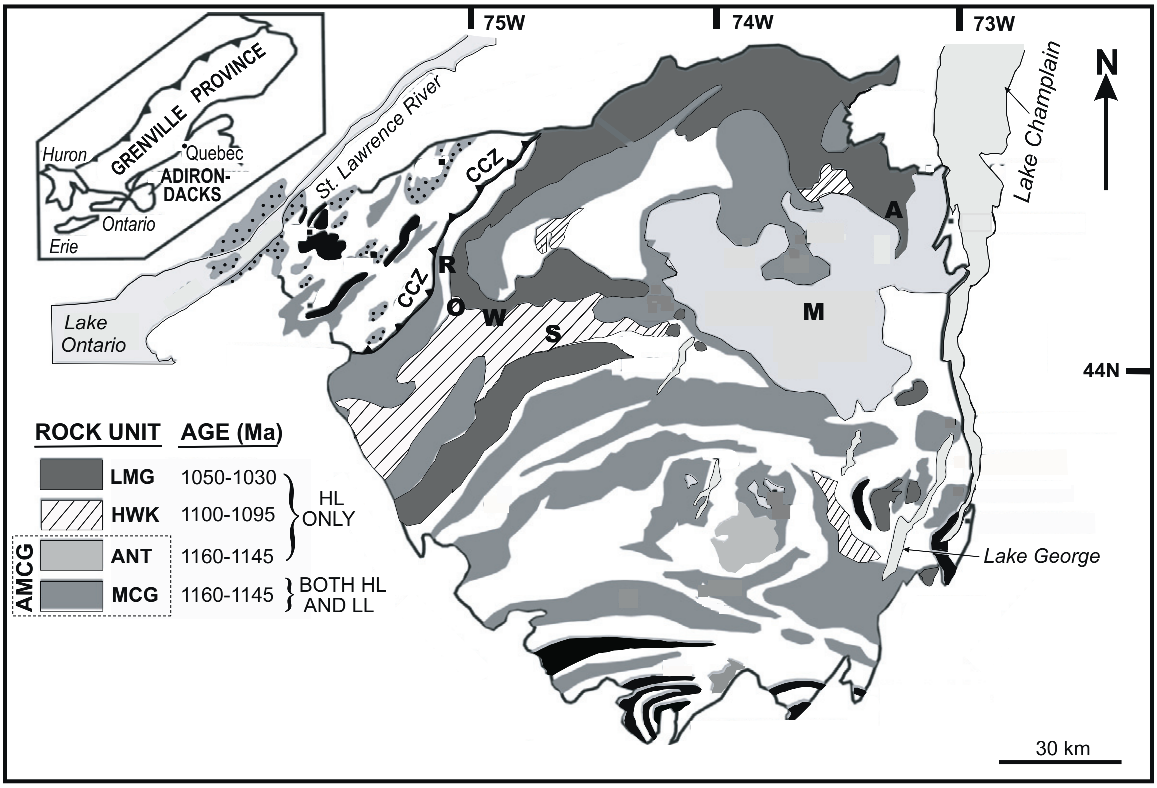

Fig. 1. Generalized geologic map of the Adirondack Mountains, New York. Inset shows the location of the Adirondacks in the southeast corner of the Grenville Province. The Adirondacks are divided in the Highlands (HL), the major part of the map, and Lowlands (LL), in the northwest corner, separated by the Carthage-Colton mylonite zone (CCZ). Predominate rock units are: the Lyons Mountain Granite (LMG), the Hawk Mountain Gneiss (HWK), rocks of the anorthosite–mangerite–charnockite–granite series (AMCG), divided into anorthosite (ANT) and mangerite–charnockite–granite rocks (MCG). Sampling locations are indicated by letters: A, Au Sable Forks, M, Marcy anorthosite, O, Oswegatchie region (OSW), R, Russell area, S, Sabattis area, W, Wanakena granite. O, R, and S locations are all in the GMS unit.
Map adapted from McLelland et al. (2004).

metamorphosed syenites and pyroxene diorites, as well as granitic gneisses.

A total of 45 sites were collected, with from 3 to 10 cores per site. Cores were either drilled with a gasoline-powered drill and oriented using a brass-orienting sleeve with magnetic and solar compasses or were drilled in the laboratory from large oriented blocks collected at the outcrops. A number of sites, mainly ones from other metamorphic rocks than the anorthosites and microcline gneisses, did not yield stable magnetizations and are not reported on here. Twenty-three sites (139 cores) provided measurable magnetizations and responded well to demagnetization. In addition to the new data on the 23 sites reported here, magnetic data for another 13 sites were provided by Rob Hargraves (per. comm., 2002) from earlier studies he instigated on the Marcy Anorthosite, associated metamorphic rocks, and the fayalite granites.

3. Magnetic mineralogy

The oxide mineralogy of the Adirondack Highlands was studied in part because of the economically valuable oxide ore deposits (Buddington, 1939, 1968; Balsley and Buddington, 1958; Buddington et al., 1963). In the rock units sampled here, the oxide mineralogy is dominated by ilmeno-hematite (hematite with ilmenite exsolution) in the metamorphosed sillimanite gneisses, and magnetite with subsidiary hemo-ilmenite in the fayalite–ferrohedenbergite granites, and metamorphosed anorthosites. A short overview of the magnetite–ulvöspinel and hematite–ilmenite series, and lamellar magnetism is given below.

3.1. Magnetite–ulvöspinel series

The cubic inverse spinel series magnetite ($Fe^{2+}Fe^{3+}{}_2O_4$)–ulvöspinel ($Fe^{2+}{}_2Ti^{4+}O_4$) forms a solid solution. Under reducing conditions a titanium-bearing magnetite will exsolve ulvöspinel on (1 0 0) cube planes with cooling. In terrestrial rocks, by oxidation–exsolution, titanomagnetite commonly forms ilmenite lamellae on (1 1 1) of magnetite, a near perfect fit with (0 0 0 1) of ilmenite (Lindsley, 1991). Magnetite, which has exsolved ulvöspinel lamellae along (1 0 0), usually shows a later oxidation of the ulvöspinel lamellae to ilmenite. 'Oxidation' may involve an external source, however it may also occur as a simple shift of tie lines with coexisting ilmenite with falling temperature. Oxidation–exsolution commonly takes place above the blocking temperature of thermo-remanent magnetization of magnetite, resulting in the magnetite having a blocking temperature that can be correlated with the composition. In metamorphic rocks and slowly cooled igneous rocks, the processes described above commonly leads to magnetite that is nearly pure $Fe^{2+}Fe^{3+}{}_2O_4$ with a 580 °C Curie temperature.

3.2. Hematite–ilmenite series and lamellar magnetism

In the hematite–ilmenite series (Fe_2O_3–$FeTiO_3$) exsolution temperatures are dependent on composition; however, cooling history will also have an effect on the amount, and size of exsolution lamellae, which in turn strongly affects the magnetic properties. The chemical behavior is dominated by two tricritical points in the 1 atmosphere phase diagram, which result in phase separation (Fig. 2). On the Fe–Ti ordering curve one point occurs at ≈Ilm

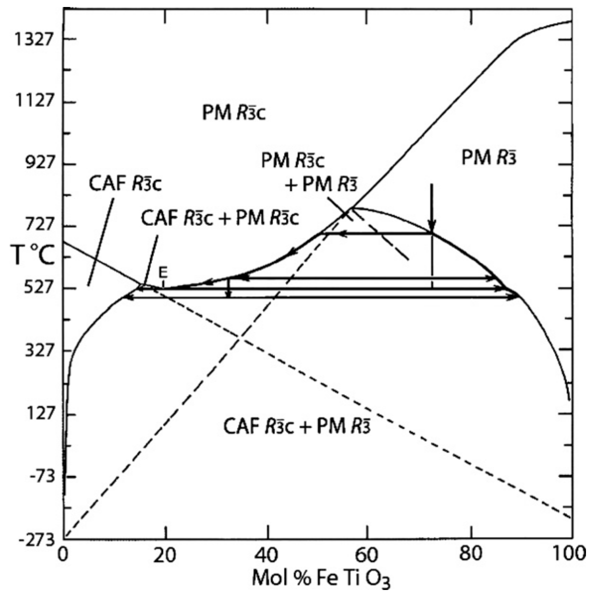

Fig. 2. Hematite–ilmenite phase diagram at 1-atmosphere showing composition–temperature relations involving Fe–Ti ordering (long-dash line) and magnetic ordering (short dashed line), miscibility gaps and two 'tricritical points'. One tricritical point lies on the Fe–Ti ordering curve at ~777 °C and 57 mol% $FeTiO_3$ creating a two-phase region of disordered paramagnetic PM $R\bar{3}c$ + ordered PM $R\bar{3}$. To the left, on the hematite rich side, the second tricritical point lies on the magnetic ordering curve at ~15 mol% $FeTiO_3$ and 530 °C, with a eutectoid point (E) of ~520 °C, below which there is a two phase region of CAF $R\bar{3}c$ hematite + PM $R\bar{3}$ ilmenite. For compositions richer in ilmenite than point E, magnetization only occurs below ~520 °C, with the onset of exsolution of a CAF hematite. For compositions more hematite rich then E, magnetization (CAF) occurs above 520 °C. Arrows show multiple stages of exsolution for a bulk composition of ~ilm72.

57, 780 °C, creating a two-phase region with a hematite-richer $R\bar{3}c$ paramagnetic (PM) Fe–Ti-disordered phase, and an ilmenite-richer $R\bar{3}$ PM Fe–Ti-ordered phase. On the magnetic ordering curve at ≈Ilm 15, ~525 °C, another two-phase region occurs with a hematite-richer $R\bar{3}c$ canted anti-ferromagnetic (CAF) phase, and an ilmenite richer $R\bar{3}c$ PM phase. The two limbs of the $R\bar{3}c$ PM phase region converge at the eutectoid (≈520 °C) with the three-phase reaction $R\bar{3}c$ PM → $R\bar{3}c$ CAF hematite + $R\bar{3}c$ ilmenite. In bulk compositions more hematite-rich than the eutectoid point, hematite will magnetize at a higher temperature. In bulk compositions more ilmenite-rich than the eutectoid point, magnetization occurs below 520 °C, by a chemical reaction with exsolution of CAF hematite solid solution from a more ilmenite-rich host. Microscopic observations of exsolution lamellae from slowly cooled rocks show exsolution has taken place in several diffusion-controlled steps (McEnroe et al., 2001, 2002, 2007, 2009).

The strong and stable NRMs of rocks dominated by finely exsolved hematite–ilmenite intergrowths (McEnroe and Brown, 2000; McEnroe et al., 2001, 2002, 2009) led to investigations through Monte Carlo simulations and ionic and chemical models (Harrison and Becker, 2001; Robinson et al., 2002, 2004, 2006) and to the theory of lamellar, or chemical interface, magnetization (LM).

Phase interfaces between CAF hematite and PM ilmenite contain $Fe^{2+}Fe^{3+}$ layers that attempt to reduce ionic charge imbalance to the oxygens (Robinson et al., 2006). These are a chemical average between hematite and the Fe layers of adjacent ilmenite, and with a magnetization average intermediate between 4 and 5 μB. The chemical situation of a lamella within a unit of rhombohedral oxide dictates that the two contact layers will have the same magnetic moment of ≈4.5 μB directed exactly along the direction of the sublattice magnetizations in the (0 0 0 1) plane, counterbalanced by the magnetic moment of one hematite layer at ≈5 μB in the opposite direction, thus giving a net moment per lamella of 4 μB. The intensity of lamellar magnetism will thus be proportional to the density of exsolution lamellae and the proportion of them that are magnetically in-phase (see McCammon et al., 2009).

LM has been investigated further theoretically with density functional simulations of the magnetic properties of lamellar interfaces (Pentcheva and Nabi, 2008), with neutron powder diffraction studies (Harrison et al., 2010) and by room and low-temperature Mössbauer (Dyar et al., 2004; Frandsen et al., 2007; McEnroe et al., 2007; McCammon et al., 2009). The experimental proof that the NRM originates from nanoscale lamellae (Fabian et al., 2008) is based on the discovery of giant exchange bias below 57 K in ilmeno-hematite rocks (McEnroe et al., 2007; Harrison et al., 2007; Fabian et al., 2008). Exchange bias requires the magnetic moment to be linked to exchange-coupled interfaces that occur between the host and the lamellae. Numerous samples discussed here have the bulk of the NRM carried by LM. One of the major differences between a rock with magnetite, which has a thermoremanent magnetization (TRM) and one with exsolved hematite–ilmenite with LM, is that unblocking temperature of LM in short term laboratory experiments, is much higher than the original acquisition temperature. LM is a chemical remanence (CRM) acquired during the exsolution process, and not a TRM, even though the exsolution process is thermally controlled. In general the unblocking of multidomain (MD) magnetite is higher than the acquisition temperature. However, here the difference is that the remanence carried by MD magnetite is of dubious geological significance. This is why Schmidt (1993) strongly advocated eradication of MD remanence in paleomagnetic and paleointensity studies using low temperature demagnetization employing liquid nitrogen.

Observations of exsolution microstructures are necessary in both the magnetite–ulvöspinel and hematite–ilmenite series to characterize the remanence carriers. To estimate the times of acquisition of the stable magnetization, it is necessary to understand the temperatures at which exsolution occurred based on phase diagrams, and the timing of the exsolution process with respect to minerals used for determining the geochronology of the cooling process, in particular titanite, rutile and hornblende following peak metamorphism.

3.3. Magnetic mineralogy of metamorphosed rocks and fayalite granites

McEnroe and Brown (2000) characterized mineral chemistry and oxide mineralogy of rocks of the Russell Belt, a part of the GMS unit, in a detailed study correlating the oxide mineralogy, magnetic properties and local magnetic anomalies. Kasama et al. (2004) discussed exsolution microstructures in the ilmeno-hematite grains from the GMS unit. These grains have lamellae of ilmenite, ±rutile, spinel and pyrophanite. Exsolution lamellae range in size from a few microns down to a few nanometers. While magnetite is largely absent in the Russell GMS samples, whereas in the Oswegatchie (OSW) 1–3% multidomain magnetite is observed in thin section which is in good agreement with estimates from measured susceptibility values. Ilmeno-hematite samples examined by TEM showed exsolution lamellae ranging in thickness from microns to 1 nm, where the unit cell of ilmenite is 1.4 nm thick. Kasama et al. (2004) demonstrated a strong correlation between the amount of exsolution in the oxides and the intensity of the NRM. Samples with ilmeno-hematite with abundant exsolution lamellae had NRM values of 5–8 A/m, whereas the sample with titanohematite and only rare exsolution, mostly of rutile lamellae, had an NRM < 0.2 A/m.

A full discussion of the acquisition of magnetization is presented later. Here new petrographic observations on the oxides from stable paleomagnetic sites in metamorphic rocks, the fayalite granites,

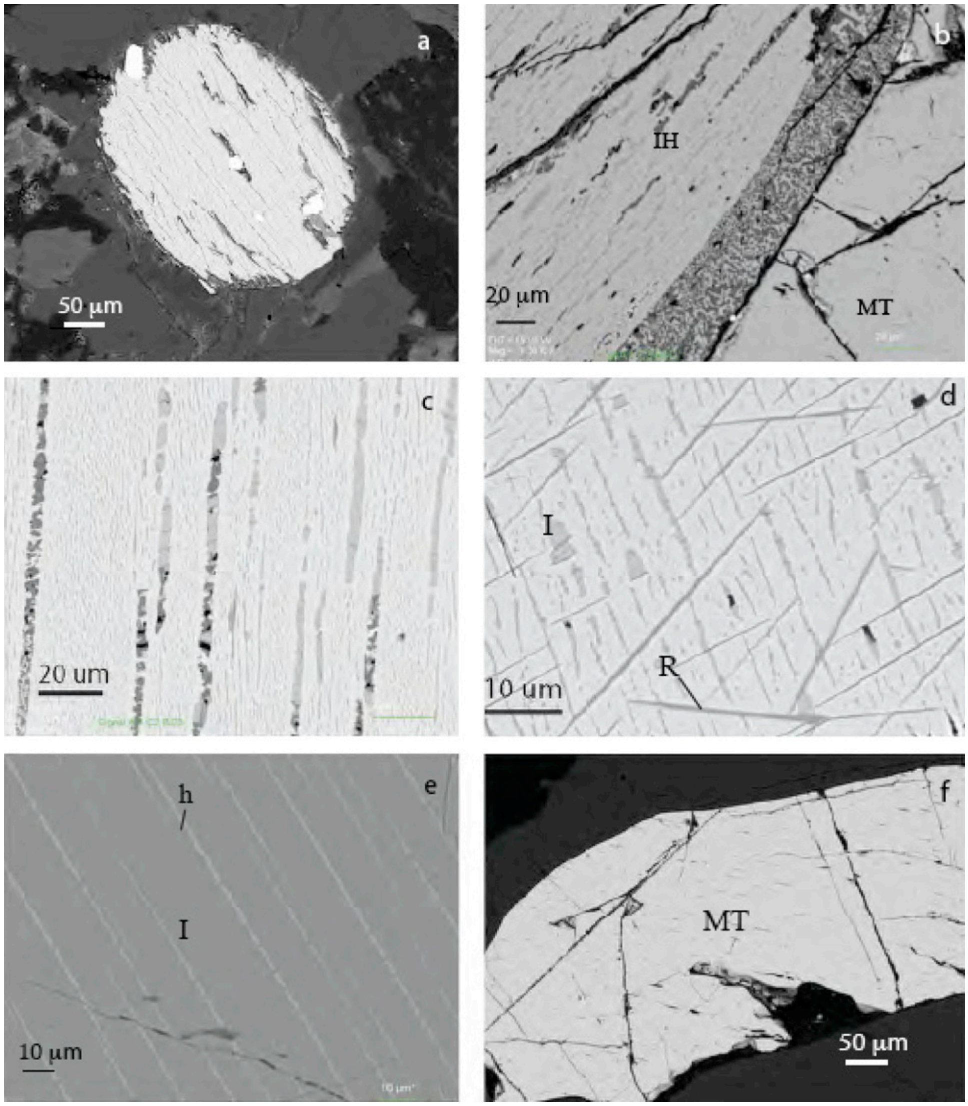

Fig. 3. Scanning electron backscatter images: (a) Discrete ilmeno-hematite grain with ilmenite exsolution on (0 0 1) and spinel (black). Bright white inclusions are monazite grains. Scale bar 50 μm. (b) Ilmeno-hematite with lamellae of ilmenite in a titanohematite host. A symplectite of hematite (white), rutile (gray), and spinel (black) is at the border of a large magnetite grain. Scale bar 20 μm. (c) Ilmeno-hematite (IH) with two generations of ilmenite exsolution (gray) from Sabattis. Large ilmenite lamellae contain rutile blades. Some 1st generation ilmenite lamellae also contain pyrophanite (light gray) + rutile + spinel (black). Scale bar 20 μm. (d) Ilmeno-hematite grain of titanohematite host with rutile needles, ilmenite and spinel. Minor spinel and rutile are associated with 1st generation ilmenite lamella. Scale bar is 10 μm. (e) Hemo-ilmenite. Ilmenite (I) host grain with hematite lamellae (H), which are widely spaced and only two generations are observed. Scale bar 10 μm. (f) Large magnetite grain with rare spinel and oxidation exsolution of ilmenite. Spinel blades commonly form at the rim of ilmenite lamellae Scale bar is 50 μm in length.

and the anorthosites based on reflected-light microscopy and SEM observation is given.

3.3.1. GMS unit

The GMS unit and associated rocks have an oxide mineralogy dominated by the Fe_2O_3–$FeTiO_3$ system. Millimeter size ilmeno-hematite grains (titanohematite host with ilmenite exsolution lamellae) are found throughout this unit. Exsolution lamellae of ilmenite ($FeTiO_3$) are ubiquitous and some samples also contain pyrophanite ($MnTiO_3$) parallel to (0 0 0 1) plane of titanohematite. Rutile, spinel and corundum (Fig. 3a-d) are also common. Multiple generations of exsolution are common and range from the micron scale down to the nm-scale. In the Russell area (AD 1–6 and 34) ilmeno-hematite is the dominant oxide and magnetite is either absent, or very rare. In the few samples that contain magnetite, it is generally present as relict grains that are inside secondary maghemite. The maghemite formed by replacement of magnetite by 'martitization'. Less common are near end-member hematite grains with minor ilmenite exsolution and abundant rutile needles. McEnroe and Brown (2000) and Kasama et al. (2004) discuss these oxide assemblages in detail.

In the OSW area ilmeno-hematite and multi-domain magnetite occur together. The magnetite grains may contain minor oxidation–exsolution lamellae of ilmenite, but commonly the magnetite grains are free of ilmenite lamellae (Figs. 3b and 4a). The ilmeno-hematite is full of oriented ilmenite lamellae, usually less than a micron wide, and a few larger lamellae, which appear to be further altered to a mixture of ilmenite, rutile, spinel and hematite (Fig. 3c). When ilmeno-hematite and magnetite are in contact typically a symplectite of spinel, rutile and titanohematite (Figs. 3b and 4a) is present. Commonly rutile is observed as plates parallel to the octahedral planes of the titanohematite and is intimately associated with ilmenite lamellae parallel to the (0 0 1) planes (Fig. 3d).

The oxides from samples in the Sabattis area (sites AD 15 and 18) are very similar to those in other parts of the GMS unit, with

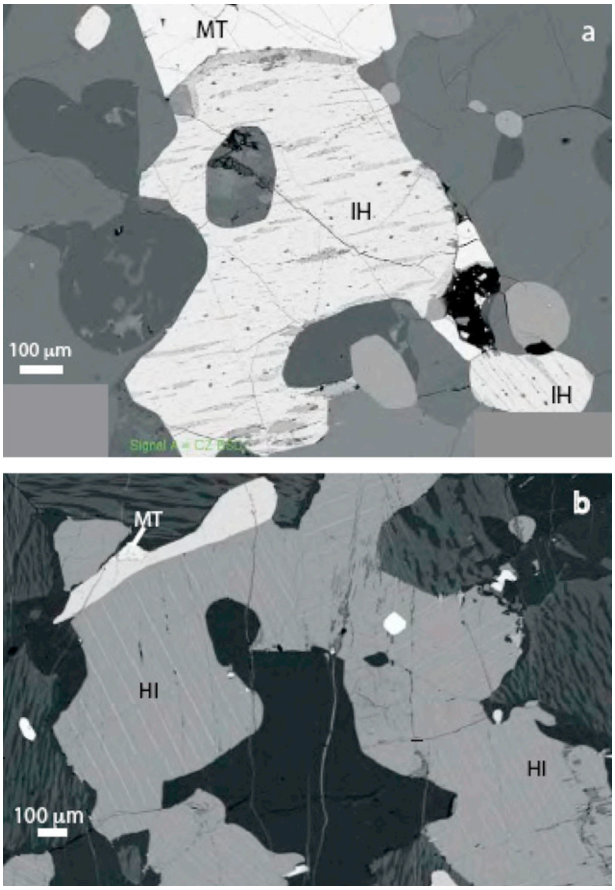

Fig. 4. Scanning electron backscatter images: (a) Oswegatchie sample showing large grains of ilmeno-hematite (IH) and magnetite (MT). Exsolution in ilmeno-hematite is same as in Fig. 3b. Magnetite has minor spinel and rare oxy-exsolution of ilmenite. (b) Minor hemo-ilmenite (HI) and magnetite (MT) in sample from Wanakena granite. Large ilmenite grains are rare, and contain only minor exsolution lamellae of hematite. Magnetite grains are generally free of lamellae. Scale bar in both images is 100 µm in length.

large, generally millimeter size ilmeno-hematite grains. In some samples magnetite appears to be absent (site 18) while in AD 15 it composes almost 0.5% of the rock. Large rutile plates (>50 µm long) are found throughout the grains except at the rims (Fig. 3d). Though the oxides in the Sabattis area are similar in character to the others GMS rocks, they are less abundant in thin section.

3.3.2. Metamorphosed anorthosites

The anorthosite and gabbroic anorthosite samples vary in abundance of oxides though generally the oxides compose ~0.5% of the rock. Numerous samples contain coarse multi-domain (>20 µm) magnetite grains typically 0.5–1.0 mm in diameter, and rare pseudo-single-domain magnetite (Fig. 5c). Large magnetite grains have spinel needles, and minor oxy-exsolution lamellae of ilmenite (Fig. 3f). Spinel rods, or needles, commonly form at the edges of the ilmenite lamellae. Numerous anorthosites contain hemo-ilmenite in addition to magnetite. The hematite lamellae are minor and typically less than a micron thick. Some grains show alteration, with ilmenite altered to a rutile + hematite. Other anorthosite samples have magnetite as described above and ferrian ilmenite without exsolution lamellae of hematite. In some anorthosite samples, hemo-ilmenite and magnetite–ilmenite 'inclusions' are well preserved (Fig. 5a–c) within large plagioclase grains. Hemo-ilmenite inclusions commonly show two generations of exsolution (Fig. 5a and b) and magnetite has oxidation–exsolution of ilmenite

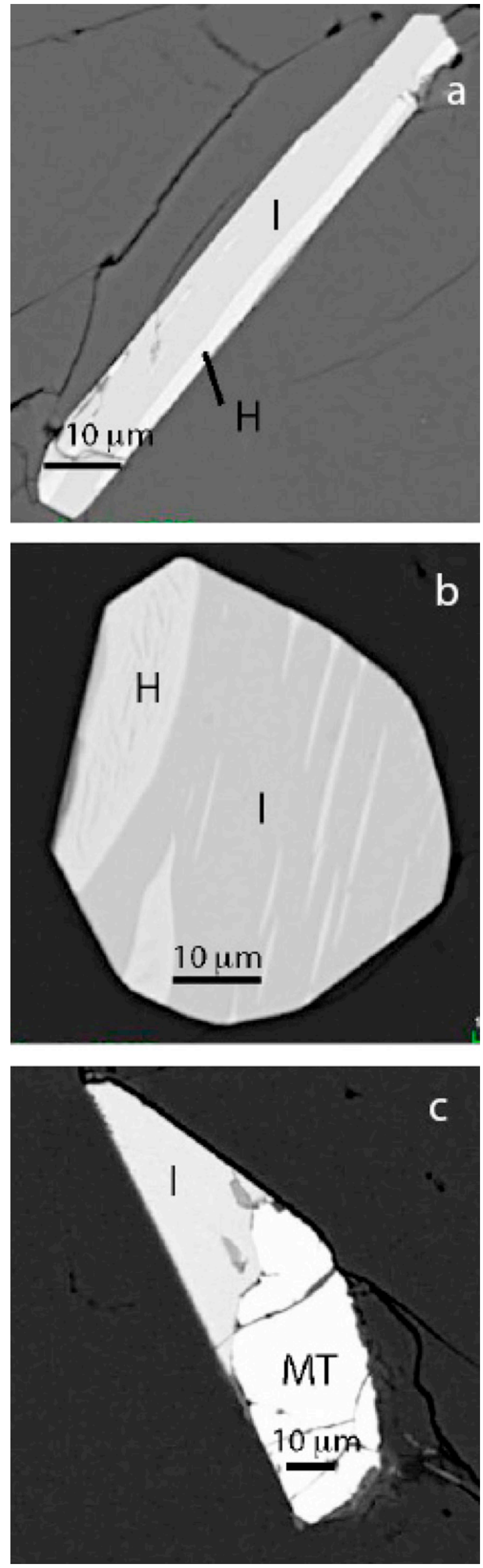

Fig. 5. Scanning electron backscatter images of oxide inclusions in meta-anorthosite samples. (a) Hemo-ilmenite inclusion in plagioclase grain. Two generations of hematite lamellae (H-white) in and ilmenite host (I). (b) Hemo-ilmenite grain with multiple generations of hematite lamella in ilmenite host (I). Note large hematite lamellae (H) have abundant smaller ilmenite lamellae. (c) Magnetite (MT) inclusion in plagioclase; the magnetite shows oxidation–exsolution lamellae of ilmenite (I). All scale bars are 10 µm in length.

lamellae. Pyrite is a common accessory phase. Overall the anorthosites are significantly more reduced than the GMS rocks, probably due to original more reduced bulk composition of the rocks before metamorphism.

3.3.3. Wanakena granite

Magnetite and hemo-ilmenite are the dominant Fe-oxides in the Wanakena Granite and compose approximately 0.5% of the rock. Magnetite is present as minor discrete grains, in multi-domain and pseudo-single domain sizes. Magnetite has oxidation–exsolution lamellae of ilmenite and spinel. Hemo-ilmenite (ilmenite with hematite exsolution) grains are up to a few millimeters in size and contain fine hematite lamellae, commonly less than 1 μm thick (Fig. 4b). Two generations of lamellae are observed at the optical scale. Minor oxide exsolution occurs in the feldspars. Ilmeno-hematite (hematite with exsolution of ilmenite lamellae) has not been observed in any thin sections from the Wanakena Granite indicating that it is less oxidized than the GMS unit.

4. Paleomagnetism

4.1. Laboratory procedures

Paleomagnetic directions were measured on a 2G SQUID magnetometer at the University of Massachusetts, Amherst. Characteristic directions were determined after progressive alternating-field (AF) and thermal demagnetization procedures using principal component analysis techniques described by Kirschvink (1980). AF demagnetization was performed in steps up to 100 mT using a Molspin AF demagnetizer. Because of the very high stability to AF demagnetization, typically with median destructive fields (MDF) exceeding 100 mT, thermal demagnetization was the primary method used. Thermal demagnetization behaviors in steps up to 680 °C using an ASC thermal demagnetizer were investigated. Magnetic susceptibility values were measured on a Sapphire susceptibility bridge. Isothermal remanent magnetization measurements were made using an ASC Pulse Magnetizer with a peak field of 1.25 T. Hysteresis measurements at room temperature in fields of up to 1.5 T were made on Princeton Applied Research micromagnetometer.

4.2. Magnetic properties

4.2.1. Susceptibility and NRM

Average values of magnetic susceptibility and natural remanent magnetization (NRM) for 23 sites from the Adirondack rocks are listed in Table 1 and represented in Fig. 6. Samples from the GMS unit show a wide range in susceptibility and NRM values, with site mean susceptibilities ranging from 1×10^{-4} to 1×10^{-1} (SI units), and site mean NRM varying from 0.02 to 7.0 A/m. Large susceptibility variations are related to the presence or absence of magnetite. Rocks from the Russell area containing only ilmeno-hematite or hematite have susceptibilities averaging 6×10^{-3}. GMS rocks from the Oswegatchie area, which contain both ilmeno-hematite and magnetite, have susceptibility values which average 1×10^{-1}. High NRM values (averaging 5.7 A/m) are associated with all the samples from the OSW area, whereas the Russell area rocks show a wide range of NRM values, from 7.0 A/m down to 0.02 A/m. The highest NRM values are from sites with only ilmeno-hematite, and the lowest are samples with hematite that has rare exsolution lamellae, and this is commonly rutile. The Wanakena granite is homogenous unit and shows little variation in susceptibility and NRM between sites, with an average susceptibility of 6×10^{-3} SI and average NRM of 0.7 A/m. The Marcy anorthosite samples showed a wide variation in both susceptibility (0.0006–0.08 SI) and NRM (0.1–2 A/m)

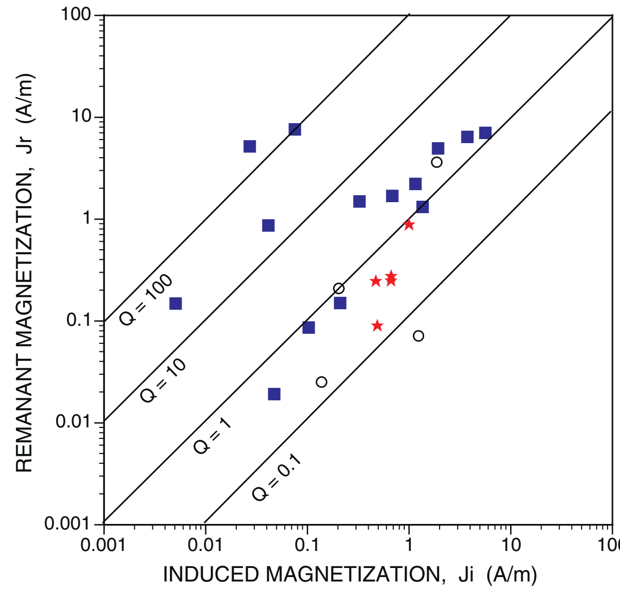

Fig. 6. Plot of induced magnetization (Ji) versus natural remanent magnetization (Jr) by sites. Diagonal lines are constant Koenigsberger (Q) values, as labeled. Solid squares, microcline gneiss sites; open circles, metamorphosed anorthosite sites; and stars, fayalite granite sites.

values, but are represented by available data from only four sites. A plot of remanent magnetization versus induced magnetization, using a local field of 46.2 A/m, is shown in Fig. 6. Constant Koenigsberger ratios (Q) of remanent to induced magnetization are marked as straight lines on the diagram, showing that over half of the sites have Q values ≥1.0. These include all the sits from the GMS that have magnetic properties controlled by ilmeno-hematite. All sites from the Wanakena granite, as well as sites from rocks dominated by magnetite, fall below the $Q = 1.0$ line.

It is also possible to consider the magnetic properties of rocks by their polarity. However this approach combines the rock types with magnetic mineralogies with different blocking temperatures (rhombohedral and cubic oxides) together and has little validity. This comparison is also limited because there are only 5 normal sites compared to 18 reverse sites. However, the NRM value for all reverse polarity sites is higher at 2.31 A/m, while the normal sites are 0.87 A/m. Susceptibility values show only a slight difference with polarity, with reverse sites averaging 0.0227 SI and normal sites averaging 0.0103 SI. This variation is simply controlled by the abundance of magnetite in the samples, reflecting the original bulk composition, and is not a time dependent feature. The OSW magnetite bearing rocks dominate the susceptibility signature of the reversed sites with the remaining sites having low susceptibilities.

4.2.2. Demagnetization studies

Alternating field studies on the GMS samples yielded extremely hard magnetizations, which were resistant to AF demagnetization, with median destructive fields commonly higher than the limit of the demagnetizer (100 mT). As shown in plots of samples from the Russell and Oswegatchie locations (Fig. 7a and c), initial strong NRM values of several A/m are not diminished in fields up to 100 mT, a behavior expected from materials having the bulk of the remanent magnetization carried by ilmeno-hematite, which is ubiquitous in these samples. Although the intensity of the NRM is slightly variable, the AF demagnetization behavior is similar in all samples. No sample from these areas loses more than 75% of the original NRM strength during, or after extensive AF demagnetization (Fig. 7a). Directions also remain constant during demagnetization,

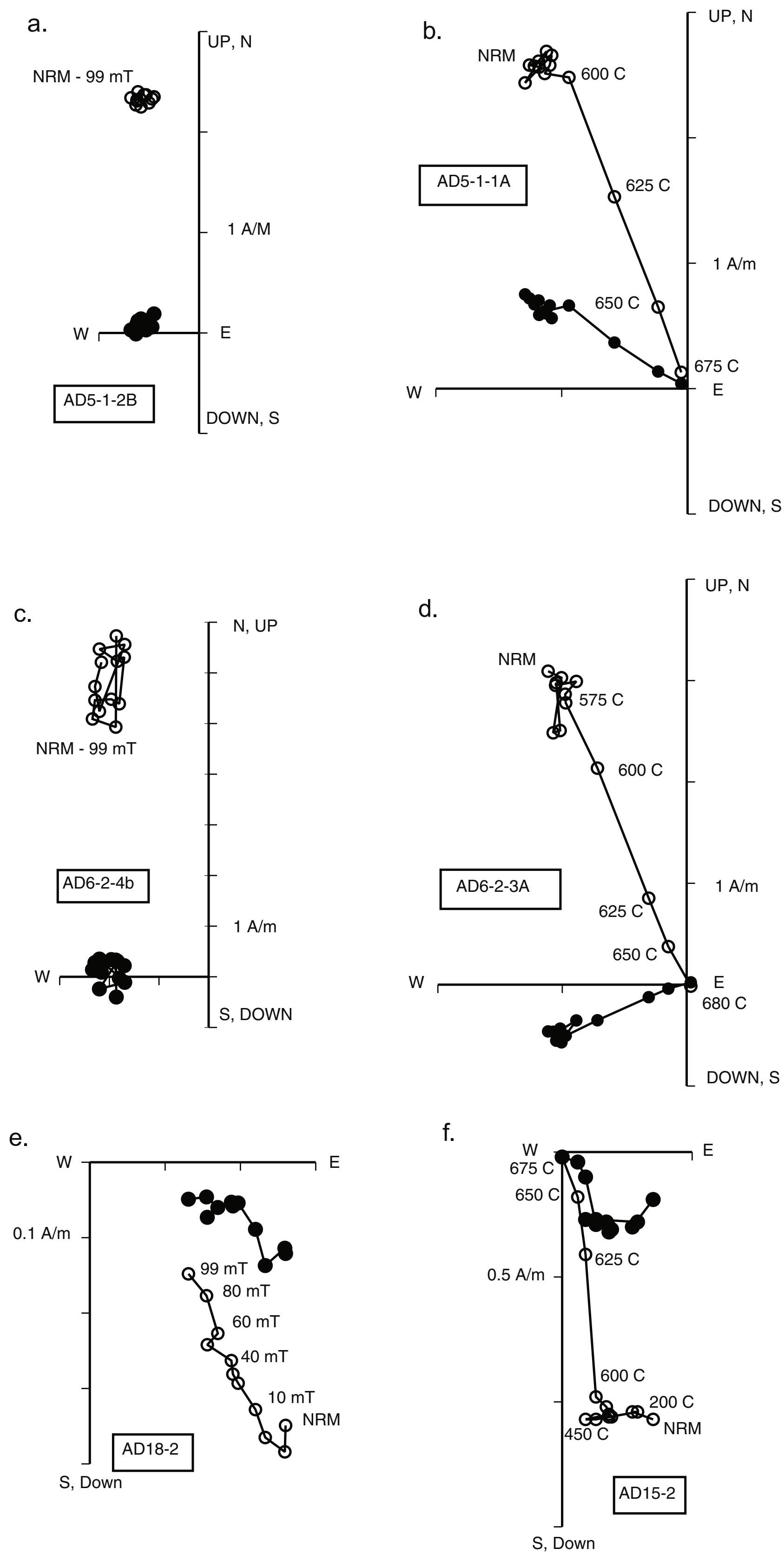

Fig. 7. Vector end-point diagrams for representative samples of the GMS rocks showing behavior with alternating field demagnetization (left column) and thermal demagnetization (right column). (a) and (b) are from the Russell area, showing no loss of magnetization on AF demagnetization, but uni-component decay after 600 °C on thermal demagnetization; (c) and (d) are from OSW area, and show similar behavior to the Russell rocks; (e) and (f) are from samples from the Sabattis area and show continuous decay with AF demagnetization but similar thermal demagnetization behavior to the other GMS samples. Open circles are vertical components; filled circles are horizontal components.

Table 1
Adirondack paleomagnetic data by sites.

Site	N/No	Susc	NRM	INC	DEC	k	α_{95}	LAT	LONG
Microcline gneiss									
AD 1	6/7	0.00059	5.16	−62.3	232.1	100	6.7	−53.1	175.8
AD 2	4/6	0.00090	0.86	−62.9	257.8	97	9.4	−36.4	164.3
AD 5	10/10	0.00707	1.49	−51.5	292.2	22	10.6	−8.0	157.4
AD 6	10/10	0.04192	4.92	−67.8	271.2	260	3.0	−31.9	152.1
AD 11	7/12	0.02508	2.21	44.8	98.1	120	5.5	12.6	349.3
AD 15	3/3	0.00457	0.15	68.9	138.7	351	6.6	12.7	308.5
AD 18	5/5	0.00224	0.09	55.3	111.6	165	6.0	11.0	334.2
AD 21	7/7	0.12257	7.02	−64.1	318.4	655	2.4	−7.1	131.8
AD 22	5/5	0.08142	6.39	−63.3	282.3	113	7.2	−22.4	152.6
AD 26	5/5	0.01485	1.69	76.9	162.9	760	2.8	19.9	291.6
AD 34	7/8	0.00163	7.61	−61.1	262.7	62	7.7	−32.3	164.4
AD 35	3/3	0.02948	1.32	−58.6	309.9	238	8.0	−4.8	140.6
AD 36	5/5	0.00011	0.15	−51.2	316.9	55	10.4	4.6	139.6
AD 40	3/3	0.00103	0.02	−55.6	294.2	130	10.9	−9.8	152.4
Metamorphosed anorthosites									
AD 12	5/5	0.08083	1.88	−61.5	304.4	44	11.7	−10.0	142.0
AD 28	5/5	0.00462	0.20	51.2	101.6	67	9.4	14.1	343.1
AD 32	3/3	0.00158	1.24	−87.5	182.2	926	4.1	−49.1	104.3
AD 33	5/5	0.00055	0.14	−65.6	290.4	153	6.2	−20.3	146.2
ADH 9	5			−75.6	312.8	210	4.3	−23.5	128.0
ADH 10	4			−76.0	279.5	177	5.3	−19.5	105.6
ADH 12	5			−73.3	240.7	798	2.2	−41.6	130.1
ADH 32	6			43.9	100.3	107	5.5	10.7	350.7
ADH 33	6			46.6	82.5	16	14.3	24.0	359.7
ADH 34	6			−73.3	240.7	49	8.2	−51.0	151.4
ADH 35	6			53.9	108.0	25	11.5	11.9	339.1
Associated metamorphic rocks									
ADH 8	4			−75.6	283.1	117	5.3	−33.4	138.0
ADH18	4			72.8	146.3	33	12.2	16.5	301.9
ADH19	4			−68.9	264.9	72	8.2	−36.5	154.7
Post-metamorphic fayalite granites									
AD 9	7/7	0.00616	0.69	−75.6	271.8	289	3.6	−37.4	139.1
AD 10	4/4	0.01989	1.04	−78.3	269.7	2409	1.9	−40.0	133.9
AD 20	4/4	0.00553	0.69	−76.4	317.3	2157	2.0	−23.3	122.7
AD 37	5/5	0.00552	0.49	−74.5	291.0	99	7.7	−28.9	135.1
AD 38	7/7	0.00201	0.51	−69.9	320.7	244	3.9	−13.4	126.6
ADH 4	4			−75.7	286.3	847	2.4	−31.9	136.0
ADH 5	4			−78.7	309.8	1406	1.9	−28.4	124.1
ADH 20	5			71.7	118.7	177	13.7	23.2	318.1

N/No, number of samples used in mean/number measured; Susc is magnetic susceptibility in SI units; NRM is natural remanent magnetization in A/m; INC and DEC are magnetic inclination and declination, in degrees; k and α_{95} are precision parameter and radius of 95% confidence circle about the mean direction (Fisher, 1953); LAT and LONG are position of virtual geomagnetic pole, in degrees latitude and east longitude. ADH sites are from R. Hargraves (per. comm., 2002).

maintaining steep negative inclinations and westerly declinations. Samples from the Sabattis region do show a loss of magnetization on AF demagnetization (Figs. 7e and 8a), with 40–50% of the NRM remaining after demagnetization to 100 mT. The demagnetized directions show steep positive inclinations and easterly declinations. These directions are nearly anti-parallel to those from the Russell and OSW areas.

Thermal demagnetization on all GMS samples showed consistent stable directions with little or no loss of intensity to temperatures above the Curie temperature for magnetite, 580 °C (Fig. 7b, d and f). Most samples show some decay between 600 and 625 °C (Fig. 8b), with half of the magnetization is lost by 625 °C, after which linear decay to the origin proceeds to 675 °C or 680 °C. Here the bulk of the stable magnetic memory is carried by ilmeno-hematite. Directions are constant during the heating steps, with steep negative directions in Russell and OSW, and steep positive directions in Sabattis samples.

The anorthosite samples are the only samples in this study to show distinctive secondary components (Fig. 9a), a steep normal component is removed by 5–10 mT. Decay after this level is slower and directions retain steep negative inclinations and westerly declinations, similar to those found in the GMS samples. Overprint directions close to the present field are also observed in the thermal demagnetization (Fig. 9b); these are removed by 400–525 °C, followed by a rapid decay to the origin, with all intensity removed by 600 °C. Demagnetization behavior, as well as the dominant loss of intensity between 550 and 575 °C (Fig. 8b) agrees well with a near end member magnetite as the magnetic carrier.

Samples from the Wanakena fayalite granite contain only one direction of magnetization and it is consistent with the neighboring GMS rocks with steep negative inclinations and westerly declination (Fig. 9c and d). Here the intensity loss is spread much more evenly over the AF demagnetization steps (Fig. 8a) and does not show the “hard” magnetism held by the Russell and OSW rocks. Inclinations, while still negative, are persistently steeper than those for Russell, OSW or the anorthosites. Thermal demagnetization on Wanakena samples yield similar directions with the intensity change beginning at low temperatures and falling off rapidly after 525–550 °C. By 575 °C less than 3% of the original magnetization remains (Fig. 8b).

4.2.3. Rock-magnetic investigations

Rock magnetic studies on samples from the different units yield results in keeping with the different magnetic mineralogy observed in each suite. Isothermal remanent magnetization (IRM) measurements are shown in Fig. 10. Two GMS samples from the Russell and Sabattis areas all show a slow increase in magnetization with increasing field strength, with saturation not occurring until fields of 0.5 T or greater. This behavior indicates the presence ilmeno-hematite in each sample, as confirmed by the

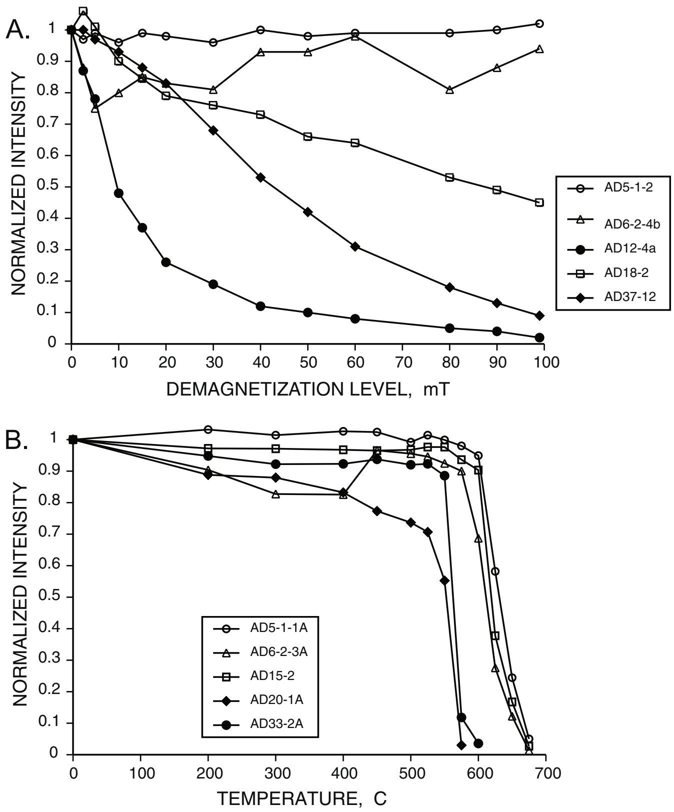

Fig. 8. Plot of normalized intensity against AF demagnetizing field level (a) and normalized intensity against temperature for thermal demagnetization (b). Open symbols are for rocks from the (GMS) microcline gneiss; closed symbols are from the fayalite granite and anorthosite rocks. Samples are also plotted in Figs. 7 and 9.

optical results. Sample AD5-2-3B from Russell belt saturates the slowest, requiring fields of at least 1 T to saturate. Sample AD1-4, also from the Russell Belt decreases in intensity from its initial condition to 0.1 T before starting to increase magnetization with full saturation above 1 T. Because it was not possible to demagnetize this sample in AF fields up to 100 mT, this sample was already partially saturated with the initial NRM of 8 A/m at the start of the IRM experiment. The NRM direction was at a large angle to its orientation in the IRM device, and the NRM needed to be rotated into the direction of the applied field before an increase in magnetization could be acquired at higher fields.

The OSW sample (AD6-2-2B) initially shows a rapid increase in magnetization until ~0.01 T where a distinct change in slope is apparent (Fig. 10) with higher fields needed to saturate the sample. This sample contains ~1% multi-domain magnetite, which saturated easily at the low applied fields, in addition to the abundant ilmeno-hematite. With higher fields the GMS samples approach saturation by 0.7 T. Samples from the Wanakena granite (AD20-2B) and the anorthosite (AD28-1B) show distinctly different behavior, both saturating at low fields, reaching full saturation by 0.2 T, as expected for samples dominated by magnetite.

Hysteresis properties were measured on 31 chips from GMS rocks, 7 samples of the Wanakena and 6 samples of anorthosite. Fig. 11 shows typical hysteresis loops representing the three rock types. The gneiss samples from the Russell area show either open loops, with coercivity values usually >150 mT from samples with only ilmeno-hematite (Fig. 11a, b and d), or slightly wasp-waisted loops (Fig. 11c), if a very small amount of magnetite is present with the ilmeno-hematite. A room-temperature hysteresis loop from an OSW gneiss sample from the Oswegatchie area (Fig. 11e) is dominated by MD magnetite. This sample contains ilmeno-hematite and >1% MD magnetite. However these samples show <25% demagnetization in AF fields of 100 mT and only minor loss in magnetization below 600 C. Detailed hysteresis properties, including hysteresis measurements with temperature are given in McEnroe and Brown (2000) as well as the arguments for ilmeno-hematite as the remanence carrier in the GMS rocks.

Wanakena fayalite granite (Fig. 11f) shows a slightly open loop more typical of PSD magnetite, or a mixture of PSD magnetite and hemo-ilmenite, both of which are observed in thin sections. The anorthosite samples show a variety of hysteresis properties that are either dominated by MD magnetite (Fig. 11g), or by PSD magnetite and hemo-ilmenite (Fig. 11h) which are present as inclusion in numerous anorthosite samples (Fig. 5c).

5. Paleomagnetic results

Paleomagnetic directions for the new 23 sites reported here are listed in Table 1, divided into three metamorphic groups (GMS, anorthosite, and associated rocks) and the post-metamorphic fayalite granite. In addition, Table 1 shows data on 13 sites from Rob Hargraves's earlier unpublished study in the 1980s and supplied to us (per. comm., 2002). These sites are from the same rock units under study here.

The metamorphic gneisses and associated rocks have dual polarity, although the NW-up directions, here called reverse (R), are predominant over the normal (N) directions (12 R and 5 N). The fayalite granite samples from Wanakena are all reversed while similar granite samples from Ausable Forks (ADH 20) are normal. The Marcy Anorthosite samples also shows dual polarity, with 4 normal sites and 7 reverse sites. There is a distinct geographic distribution of sites by polarity, with GMS and Wanakena sites to the far west (R, O and W in Fig. 1) all reversed, as are anorthosite samples from the far side of the massif just west of Lake Champlain. Normal samples of all rock types are found in the central area of the massif, including GMS samples from the Sabattis area and anorthosite samples from the western Marcy body.

Site directions are plotted in Fig. 12, with each separate rock group distinguished. Average directions for all groups, after inverting the N polarity data, are listed in Table 2. The GMS unit has average directions of $I = -62.8^\circ$ and $D = 289.2^\circ$ with $k = 29$ and $\alpha_{95} = 7.6^\circ$ for $N = 14$ sites. Anorthosite rocks give a similar direction, with $I = -65.4^\circ$ and $D = 282.9^\circ$ with $k = 25$ and $\alpha_{95} = 9.4^\circ$ for $N = 11$ sites. The average from the associated metamorphic rock group ($N = 3$) yield a steeper inclination ($I = -74.1^\circ$), but similar declination (289.5°) to the other metamorphic directions. The fayalite granites also have a steeper inclination and somewhat higher declination, with $I = -75.8^\circ$ and $D = 297.0^\circ$ with $k = 199$ and $\alpha_{95} = 3.9^\circ$ for $N = 8$ sites.

The normal and reverse directions of all the combined metamorphic rocks are quite similar, but do not pass the reversal test of McFadden and McElhinny (1990) at the 95% level. This is not surprising as we postulate the two groups (gms unit and the combined anorthosite and related metamorphic rocks) acquire their magnetization at different times as discussed in detail below. The fayalite granites have a distinct direction compared to the metamorphic rocks (Fig. 12). The inclinations of the fayalite granite are 10° steeper than the metamorphic rocks and their declinations are 10° higher. The fayalite granites do pass a reversal test, but it is at level C (McFadden and McElhinny, 1990) hampered by the presence of only one normal site. The GMS unit is represented by 10 R sites and 4 N sites; this population fails the reversal test at the 95% level although the observed angle between N and R (4.88°) is only slightly larger than the critical angle (4.68°) as determined from McFadden

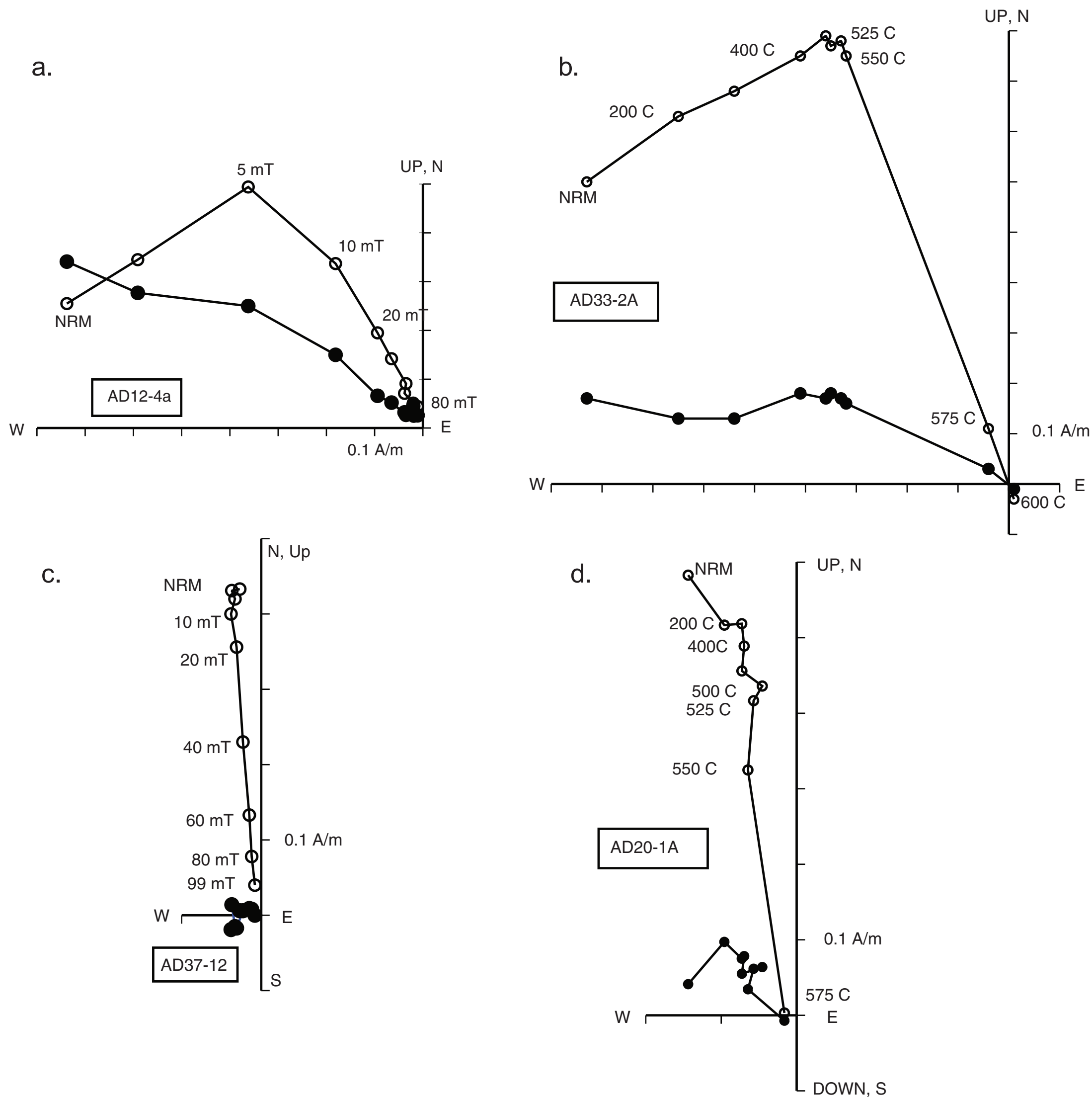

Fig. 9. Vector end-point diagrams for representative samples from the Marcy anorthosite (a) and (b), and the Wanakena fayalite granite (c) and (d). Left side shows alternating field demagnetization plots; right side shows thermal demagnetization studies on samples from the same sites. Symbols are as in Fig. 7.

and McElhinny (1990). The anorthosites and related metamorphic rocks show differences in polarity with 4 of the 5 N sites having distinctive shallower inclinations but similar declinations. The validity of averaging N and R directions is based on the small number of sites, the diverse compositions but similar metamorphic history, and similar magnetic carriers, as discussed below.

Corresponding geomagnetic poles for the various rock units are given in Table 2. The microcline gneiss unit yields a pole at 18.4°S, 151.1°E. The combination of the anorthosites and related metamorphic rocks produces a slightly steeper pole at 25.1°S and 149°E, while the post-metamorphic fayalite granites have a pole at 28.4°S and 132.7°E. The three poles are plotted in Fig. 13.

6. Discussion

A large majority of the metamorphic and post-metamorphic rocks from the Adirondacks discussed here have strong and very stable remanences, many with median destructive fields >100 mT. In most cases the samples indicate only one direction of magnetization, or in the case of overprints, as seen in the anorthosites, a secondary direction close to the present-day-field (PDF) is easily removed with AF demagnetization at low fields or thermal demagnetization by 400–525 °C. The resulting directions from both the metamorphic rocks and the post-metamorphic rocks are distinct from the present field (Fig. 12), and are also distinct from known

Table 2
Mean paleomagnetic directions by rock units.

Rock unit	N	INC	DEC	k	α_{95}	Pole	Dp/Dm
Microcline gneiss	14	−62.8	289.2	29	7.6	−18.4, 151.1	9.3, 11.8
Metamorphosed anorthosites	11	−65.4	282.9	25	9.4	−24.0, 151.6	12.3, 15.2
Other metamorphic rocks	3	−74.1	289.5	99	15.2	−30.5, 127.5	26.4, 27.9
Combined meta. rocks	14	−67.3	283.9	28	7.7	−25.1, 149.0	10.6, 12.7
Fayalite granite	8	−75.8	297.0	199	3.9	−28.4, 132.7	6.7, 7.2
All sites	36	−67.5	288.3	32	4.3	−23.2, 146.7	6.0, 7.2

Dp, Dm, semi-minor axis and semi-major axis of ellipsoid of 95% confidence about the pole; other labels as in Table 1.

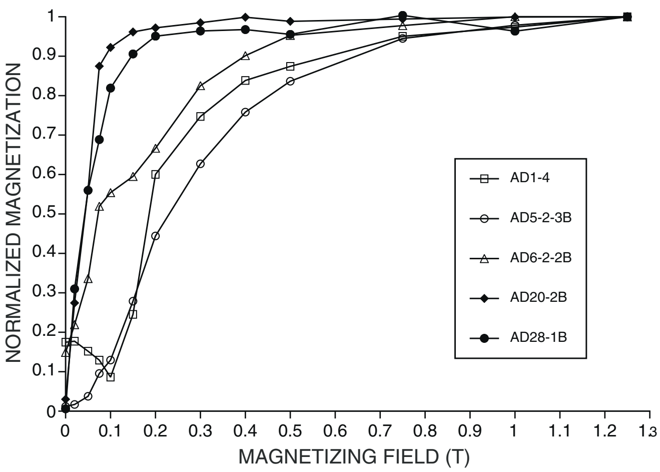

Fig. 10. Plot of isothermal remanent magnetization (IRM) normalized to magnetization at peak levels of 1.25 T. Open symbols represent samples from the microcline gneiss and indicate contributions to the magnetization from ilmeno-hematite. Close symbols are for samples from the anorthosite (solid circle) and the Wanakena granite (solid diamond) and indicate magnetite as the dominant magnetic mineral. All samples have reversed polarity except the anorthosite (AD28).

directions for Paleozoic or more recent times. Based on our petrographic observations of thin sections in transmitted and reflected light, and by SEM there is little evidence for significant chemical alteration, which might have compromised original directions since cooling from metamorphic temperatures. Widespread studies in the Adirondacks have found no evidence of later Paleozoic metamorphism, and observed mineral assemblages and structural configurations are related to Grenville activity (McLelland et al., 1996, 2010).

There is always the possibility of later tectonic and structural events that may rotate or tilt blocks of crust and thus change the orientation of the paleomagnetic signal. Extensive work in the Kapuskasing structural zone in north central Canada is such a case (Bates and Halls, 1990; Symons et al., 1994), where Archean paleomagnetic signals are tilted and rotated by later orogenic activity. The Adirondack Highlands however represent the last activity (Ottawan stage) in the lengthy Grenville Orogeny (Hynes and Rivers, 2010), and have suffered no later deformations. The Highlands, with high metamorphic grade, are separated from the lower metamorphic realm of the Lowlands by the Carthage-Colton Shear Zone (Mezger et al., 1992). The Highlands themselves are only ~125 km across, and show no evidence of separate structural regions, nor extensive faulting that would be required to produce block rotations or tilting (McLelland et al., 2010).

We assume that the magnetic directions derived from the metamorphic Adirondack rocks are characteristic directions, acquired during cooling from high-grade metamorphism ending around 1050 Ma, and after solid-state deformation associated with orogenic activity has taken place. For these rocks their original primary magnetization was completely reset during the Ottawan orogeny. The Wanakena granite possesses a primary direction acquired during cooling after its post-orogenic intrusion. Normal and reversed polarities are observed in all of the three different rock groups. In addition there is a clear geographic distribution to the normal polarity and reversed polarity sites with normal polarity sites dominantly from the central Highlands region, and the reversed sites from both the eastern and western parts of the Highlands, which we related to the cooling history. Magnetic field reversals occurred during the tens of millions of years in which the Highlands were cooling, with rocks in different parts of the orogen cooling through acquisition temperatures (~570–450 °C) at different times in different localities.

6.1. Age of remanence

Determination of the age of remanence of the rocks studied here depends on the rate of cooling from peak metamorphism, and the temperature when the magnetization is acquired. For SD and PSD magnetite this is the blocking temperature, ~570 °C; the acquisition temperatures of the magnetization by lamellae formation in the ilmeno-hematite and hemo-ilmenite samples is more complicated and is discussed below.

The Ottawan event in the Adirondacks reached its peak by 1050 Ma (McLelland et al., 2004, 2010) with temperatures greater than 750 °C and pressures of 7–8 kbar (Bohlen et al., 1985; Jaffe et al., 1978). Any previous magnetization in the metamorphic rocks would have been obliterated at this time, and new magnetizations gained as the region cooled after peak metamorphism. Numerous scenarios have been proposed for the cooling history of the Adirondacks using mineral ages and closure temperatures. Based on sphene geochronology, Mezger et al. (1991) suggest that after 1050 Ma there was slow cooling of the massif at a rate of 1.5 °C/my. At this rate, under a steady cooling regime, the rocks would reach temperatures of ~570 °C by 930 Ma. McLelland et al. (2001) suggest that the orogenic event, occurring as a Himalayan type collision, was followed by uplift and subsequent collapse of the interior of the orogen, resulting in initial rapid cooling, upwards to 20 °C/my, followed by a prolong period of slower cooling.

Several authors have produced specific cooling curves for different parts of the Adirondack area. The commonly cited peak metamorphic temperature of 750 °C is projected for the central Adirondack Highlands only, with lower peak temperatures of 650 °C found in the western part of the region (Bohlen et al., 1985). Using different peak metamorphism temperatures and mineral ages from garnet, monazite, sphene, hornblende and biotite (Mezger et al., 1991), Mezger et al. (1992) produce four different cooling curves for the central, western, and southern Highlands, and the Lowlands. Streepey et al. (2000, 2002), working along the CCSZ between the western Highlands and the Lowlands, provide additional hornblende and biotite ages from either side of the shear zone, including one hornblende age of 950 my in the region of the sampled GMS sites. The resulting cooling curves are similar but less detailed than the curves of Mezger et al. (1992). Both interpretations have cooling rates of 1–3 °C/my, and temperatures of ~500 °C reached at ~960–950 Ma.

Based on the ilmenite–hematite 1-atmosphere phase diagram (Fig. 2), exsolution from the single phase titanohematite grains to ilmeno-hematite within the GMS unit would have started just below 520 °C for bulk compositions more Ti-rich than hem_{82} (Ilm_{18}). Electron microprobe (EMP) data of GMS samples showed that most bulk titanohematite compositions were more ilmenite rich than hem_{82} (McEnroe and Brown, 2000). This estimate is based on EMP data of the titanohematite host areas, which did include overlap analyses of small lamellae, but excluded the larger ilmenite lamellae. If these lamellae are integrated into the titanohematite analyses the 'original bulk composition' will be more Ti rich than hem_{82}. Kasama et al. (2004) analyzed lamellae and hosts by TEM, which has a much smaller spot size (nm-size) than that used by the microprobe, but with lower precision. Their data shows a range in titanohematite compositions from hem_{93-84}, with a few analyses of hem_{100}. The pure hematite analyses are typically from areas in titanohematite grains where there are abundant rutile needles (see Fig. 3d) in addition to ilmenite lamellae, or from hematite grains with dominantly rutile needles and few ilmenite lamellae.

A bulk composition with less than ~15 mol% $FeTiO_3$, (or $MnTiO_3$), would have magnetized first as a canted antiferromagnetic hematite (CAF) before exsolution occurred. If exsolution had not occurred, the magnetic blocking temperature would be significantly lower for the ilmenite-richer compositions, because the

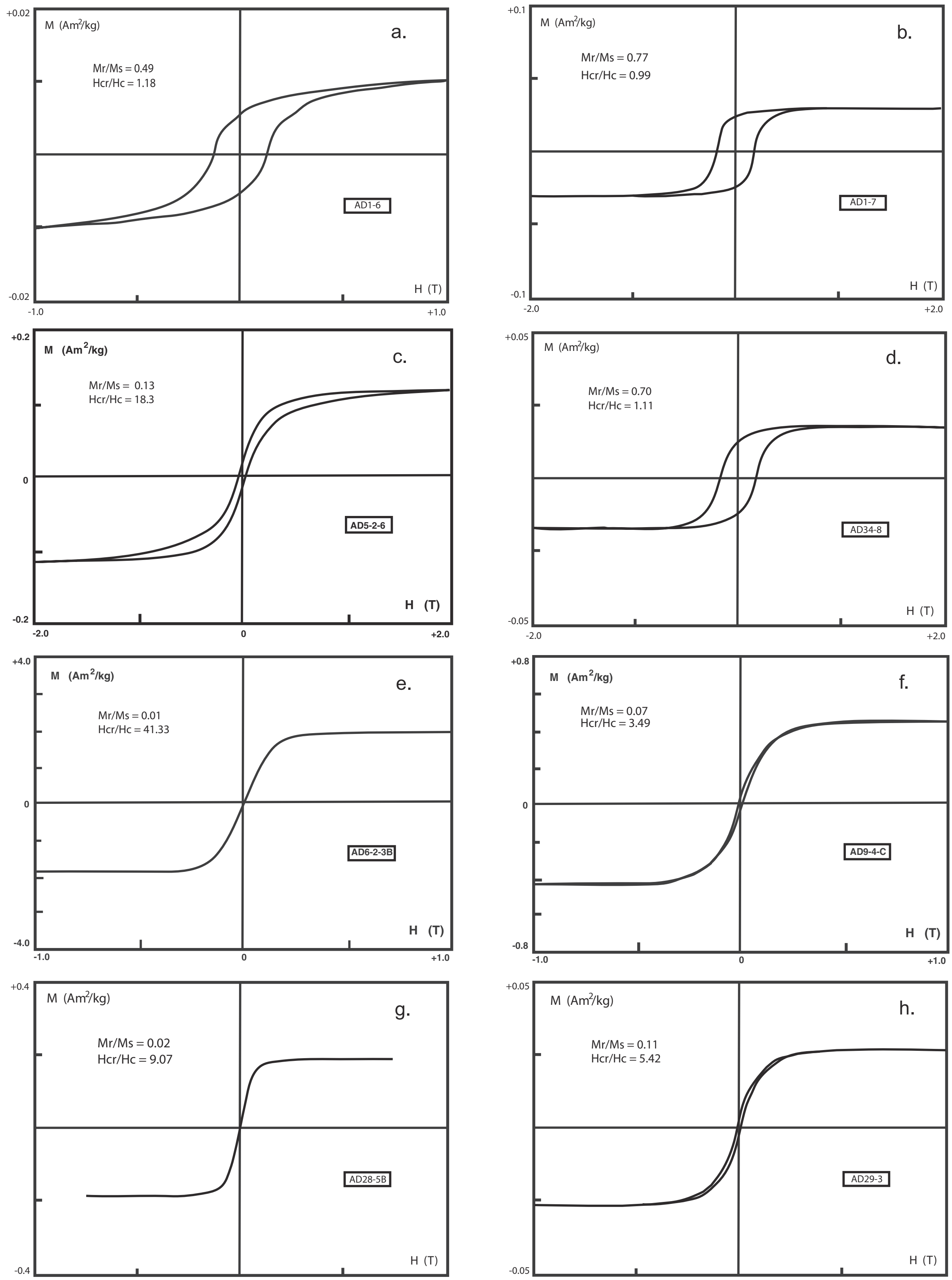

Fig. 11. Hysteresis loops, showing magnetization (Am^2/kg) resulting from applied fields up to 1.0 or 2.0 T. (a, b and d) Fat loops from samples of the Russell Belt area, where rocks only contain ilmeno-hematite, (c) wasp-waisted loop, also from Russell Belt, containing a small amount of magnetite and ilmeno-hematite, (e) sample from the Oswegatchie area containing both ilmeno-hematite and magnetite, (f) sample from the Wanakena granite indicating PSD magnetite as the dominant oxide, (g) and (h) two samples from the Marcy Anorthosite with MD and PSD magnetite.

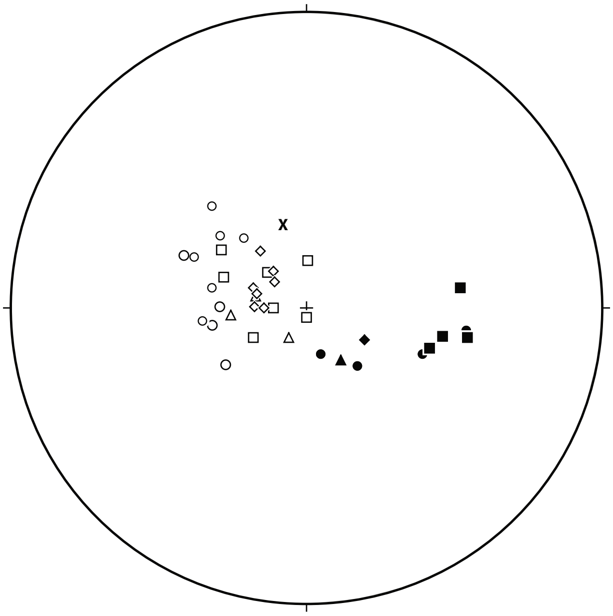

Fig. 12. Equal area stereographic net of site mean paleomagnetic directions. Open symbols are vectors on the upper hemisphere; closed symbols are on the lower hemisphere. Circles, microcline gneiss sites; squares, metamorphosed anorthosites; diamonds, fayalite granites; triangles, other metamorphic rocks. Site data listed in Table 1. X is present magnetic field position for the Adirondack Mountains.

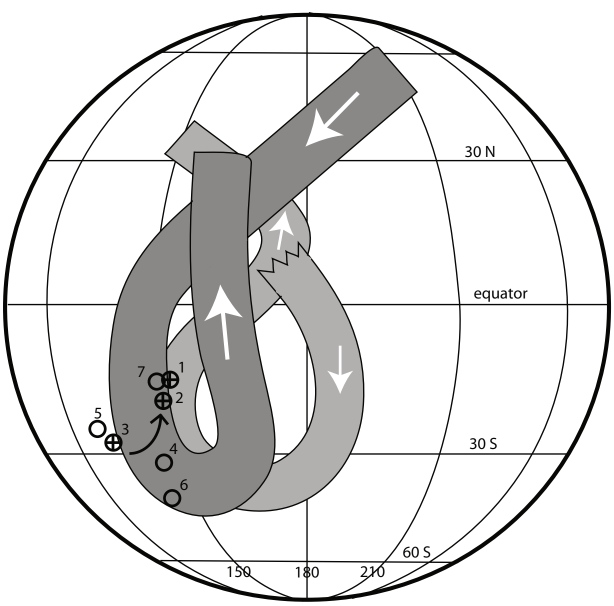

Fig. 13. Geographic plot of suggested apparent polar wander paths for Laurentia in the Proterozoic (~1100 to 740 Ma). Light grey path (clockwise motion) reproduced from Weil et al. (2004) using poles from Hyodo and Dunlop (1993) and Alvarez and Dunlop (1998); dark grey path (counter-clockwise motion) from Weil et al. (2006). Magnetic poles from this study are shown as crossed circles; open circles are other poles from Grenville rocks, as discussed in the text. Numbers refer to studies identified in Table 3. Arrows indicates direction of motion along the apparent polar wander path.

formation of lamellae drives the bulk titanohematite composition towards a purer hematite as ilmenite, pyrophanite and rutile lamellae were exsolved. For a bulk composition of hem_{60} (ilm_{40}) the exsolution temperature to obtain a CAF hematite would be 520 °C, and with the onset of magnetic ordering related to the lamellae formation with the 1st generation lamellae acquiring a magnetization just below this temperature. In comparison the same bulk composition preserved as metastable without exsolution would not magnetize until it reached an antiferromagnetic (CAF) ordering temperature of ~330 °C. The majority of the grains analyzed indicated that the titanohematite bulk compositions had maximum solution temperatures of ~520 °C. Samples richer in hematite component than the Eutectoid (E) point (Fig. 2) may acquire an NRM at slightly higher temperatures, however the bulk compositions discussed here much richer in $FeTiO_3$ than the eutectoid composition. Published ilmenite–hematite phase diagrams based on experimental and theoretical considerations (Burton, 1985; Harrison, 2006; Ghiorso and Evans, 2008) all show exsolution of a CAF to the right of eutectoid point (Fig. 2), that would have started at ~520 °C.

After formation the lamellae and host show a very high magnetic stability with respect to both temperature and viscous magnetization effects. McEnroe et al. (2004) discuss experimental work on the stability of the lamellae at temperature and pressures to 10 kbar.

TEM observations (Kasama et al., 2004) show numerous generations of lamellae, with larger ilmenite lamellae containing smaller hematite lamellae. At the interfaces of the larger lamellae there are dislocations between the titanohematite host and ilmenite lamellae. Dislocations are absent at the interfaces of the abundant small lamellae and host and these show significant strain shadows. These features result in high magnetic stability and are common in rocks that have exsolution lamellae in the ilmenite–hematite system from slowly cooled rock bodies (McEnroe et al., 2001, 2002, 2007, 2009). The large range in sizes of exsolution lamellae indicate that this process continued for a long time, likely until diffusion of titanium from titanohematite ceased. By studying the range of exsolution lamellae McEnroe et al. (2002) concluded that the last set of very fine lamellae (a few nm in size) resulted from super saturation and precipitation. One of the challenges in predicting the 'magnetic closure temperature' in the ilmenite–hematite solid solution system is that exsolution could proceed in several steps over an interval of ~100 °C. With a slowing rate magnetization could be acquired over a long time span. If the bulk of the lamellar magnetism is carried by the finest lamellae, as seems likely, then most of the remanence will reflect the end of such a exsolution process and the younger end of the cooling would be most applicable. Based on the TEM compositions and the 1-atmosphere phase diagram exsolution would have continued in the GMS rocks until ~420 °C. Using the faster cooling rate this would yield a minimum 'closure temperature for the magnetization age' of around 940 Ma.

The metamorphosed anorthosite, related metamorphic rock, and fayalite granite samples are all dominated by magnetite. The magnetite is near end-member and magnetic blocking would have occurred near 570–560 °C. Though some magnetite grains do show oxidation–exsolution of ilmenite, this process would likely have occurred above 580 °C. The anorthosite and fayalite samples are significantly less oxidized than the GMS unit and the co-existing rhombohedral oxide, when present, is a hemo-ilmenite. Exsolution of hematite lamellae from an ilmenite with a bulk composition of ~ilm 90 would occur below 490 °C. However, most of the magnetization in these rocks is carried by magnetite, with hemo-ilmenite volumetrically subordinate to magnetite.

Using the published cooling curves with remanence blocking temperature for near-pure magnetite and ilmenite–hematite discussed above, we can determine approximate ages for remanence acquisition. The first rock type to lock in remanence is the fayalite granite with magnetite the dominant oxide, which reaches temperatures of 570 °C at ~990 Ma. The metamorphic rocks (anorthosites and others) located in the central Highlands have higher peak temperature and thus reach 570 °C after the granites, with a lock-in age of ~970 Ma. The GMS rocks, dominated by ilmenite–hematite, must cool to <520 °C before lamellae are formed and the remanence begins to be acquired; using 500 °C a mean temperature with sufficient remanence indicates a maximum acquisition age of ~960 Ma. The metamorphic anorthosites, and the GMS rocks have different magnetic oxides resulting in their magnetizations acquired at different temperatures. Considering the different peak temperatures and slightly different cooling paths for the two parts of the Highlands result in magnetizations acquired at slightly different times.

6.2. Laurentia pole path at Grenville time

The attempt to define a polar wander path for the North American craton during Precambrian time has a long history, dating back to early studies in metamorphic rocks of the Grenville Province (Dubois, 1962). In general poles assumed to be roughly 1.0 Ga fall in the southwest Pacific Ocean and appear to represent an APWP that forms a large loop from equatorial positions down to mid-southern latitudes and returning to equatorial poles. Dubbed the 'Grenville Loop' it represents data derived from Laurentian rocks in the age range 1100–700 Ma. Although it generally has been considered to exhibit clockwise motion (e.g. Roy, 1984; Alvarez and Dunlop, 1998), other interpretations of the data allow for counter-clockwise motion (Pesonen and Neuvonen, 1981; Weil et al., 1998). While there seems to be agreement on the older end of the loop (Keweenawan poles around 1100 Ma; Buchan et al., 2001) and Neoproterozoic poles (~800 Ma; Weil et al., 2004), the southern extent of the loop remains ambiguous.

To define a usable APWP for a given craton over a prescribed time interval it is necessary to have both well-defined stable remanent magnetizations and well-determined ages of the remanence. This becomes a difficult task when working in ancient orogenic belts with extensive metamorphic terrains, as noted in the 'key poles' criteria suggested by Buchan et al. (2000); these authors find no acceptable poles from Laurentia for the period between 1090 and 720 Ma. Although this may be an extreme limitation on the data, it is important to consider both the reliability of the remanence data, and the veracity of the age of magnetization when contemplating pole tracks.

We have considered the wealth of magnetic data published on Grenville rocks with specific attention to single component, well preserved directions observed in a number of distinct sites with numerous samples and extensive demagnetization. We also are looking for studies that are able to pinpoint the carriers of the remanence, and establish a good estimate for the age of magnetization. Many published studies suffer from limited sampling sites, poorly defined secondary components, and only cursory attention to magnetic carriers, however there remains a number of studies which are extensive and well described, producing acceptable characteristic directions.

The most difficult criteria to establish are the ages of remanence, particularly in the Grenville Province, which has a large number of distinct lithotectonic areas with different geologic histories and ages of metamorphism (Rivers, 1997, 2008). In regions of granulite-grade metamorphism where peak temperatures are >650 °C, the characteristic magnetization is acquired during orogenic cooling. One way to determine an acceptable age of magnetization employs isotopic dating of the material using a method providing a closure temperature close to the blocking temperature of the magnetic carrier. In most cases this correlation is made using $^{40}Ar/^{39}Ar$ on hornblendes yielding temperatures close to the blocking temperature of low Ti magnetite. The second method uses cooling curves constructed from several isotopic studies for one region and comparing it to the blocking temperatures of oxide phases involved. Both methods require information of the magnetic carriers, which will lead to different ages depending on magnetic phases and the initial bulk compositions of the oxides. If the only magnetic phase is a low Ti magnetite a blocking temperature of ≈560–570 °C is quite acceptable. However, if hemo-ilmenite, or ilmeno-hematite is the predominant phase, then acquisition temperatures are lower, and thus remanence blocking ages would be younger than expected for magnetite.

Investigation of the considerable number of studies on Grenville paleomagnetism leaves us with a rather small selection of results discussed here. Studies were omitted first by location. The general area of Grenville rocks includes several different metamorphic events: some areas suffer through all the metamorphic events, while other areas are "metamorphic lids" and do not undergo all the orogenic events (Rivers, 2008). As we are concerned with magnetic directions obtained in the Ottawan phase of Grenville activity, studies with earlier (Shawinigan phase) or later (Rigolet phase) metamorphisms were not used. A second important feature of acceptable studies is the number of sites and samples used in determination of a pole position. At least eight sites with an average of 5 samples per site are preferred. Finally, the age of remanence needed to be determined, either by associated radiometric dating of the orogenic event, determination of cooling curves, and consideration of the magnetic mineralogy. In some studies this is a major portion of the reported work; in others only a regional metamorphic age is mentioned.

Studies used in this compilation as listed in Table 3 and plotted in Fig. 13, along with the three separate units from this study. As can be seen in the pole plot, the three studies from the Adirondack Highlands fall in a progression from oldest (Wanakena granite) to youngest (microcline gneiss) with the metamorphosed anorthosite pole in-between. From this association it appears that the pole path for Laurentia at this time (990–960 Ma) is moving in a counter-clockwise direction as suggested by Weil et al. (1998). Results from the Haliburton Intrusions (Buchan and Dunlop, 1976; Warnock et al., 2000) in the Bancroft Terrain of the western Grenville Provence have the most detailed age determination. $^{40}Ar/^{39}Ar$ dating on hornblendes from the same outcrops as paleomagnetic samples indicate an age of remanence in these magnetite bearing rocks as 1015 ± 15 (Warnock et al., 2000), although the resulting lies between the Wanakena granite and the metamorphosed Adirondack rocks, albeit further south. The Whitestone Anorthosite and Diorite, located near Parry Sound in a region effected by the Ottawan phase, has been studied by Ueno et al. (1975) and later dated with $^{40}Ar/^{39}Ar$ studies on hornblendes by Dallmeyer and Sutter (1980) yielding an age of 980 Ma. The mineralogy reported by Ueno et al. (1975) consists of ilmeno-hematite and hemo-ilmenite with fine lamellae; thus actual acquisition of the remanence is younger than the hornblende age. The reported pole is close to the gneiss and metamorphic anorthosites, as expected for magnetization younger than 980 Ma. North of the Adirondacks is the Morin terrain, of similar geologic history, but with an earlier onset of Ottawan metamorphism (Peck and Valley, 2000). Irving et al. (1974) studied the magnetization of the Morin Anorthosite in detail and noted that the high coercivity characteristic remanence was carried by ilmenite with exsolved hematite. Following the discussion of the microcline gneiss samples presented here, the acquisition temperature for the Morin samples was probably similar, less than 520 °C. Detailed cooling curves do not exist for the Morin terrain, and the best estimate for peak metamorphism is 30–90 my earlier than in the Adirondacks (Peck and Valley, 2000) requiring the magnetization to be older than 960 Ma. Magnetic studies on the Magnetawan sediments (McWilliams and Dunlop, 1975), located in the Parry Sound area of the Grenville Province, yield a metamorphic pole position close to the Wanakena granite. Only regional metamorphic age control is available giving an average age of 950 Ma. The hornblende data cited above from Dallmeyer and Sutter (1980) suggests an older age, closer to 980 Ma for magnetization acquisition in this area.

Despite being very selective in our acceptance of pole positions and ages, it is still not a perfect fit of data. The three poles from the Adirondacks, on which we have the best age information, strongly

Table 3
Selected Grenville Province poles and age of remanence.

Study	S/s	Min	Age	INC and DEC	Pole	Reference
1. Adirondack GMS	14/89	Ilmeno-Hem	960[a]	−63, 289	18S, 151	This paper
2. Adirondack metamorphic anorthosites and other rocks	14/72	Mgt	970[a]	−67, 284	25S, 149E	This paper
3. Adirondack fayalite granite	8/40	Mgt	990[a]	−76, 297	28S, 133E	This paper
4. Haliburton intrusions	32/138	Mgt	1015 ± 15[a]	−72, 278	33S, 142E	Buchan and Dunlop (1976), Warnock et al. (2000)
5. Magnetawan metaseds	23/83	Mgt, Hem	950[b]	−73, 302	24S, 130E	McWilliams and Dunlop (1975)
6. Morin anorthosite	24/117	Hemo-Il	>960[d]	−77, 266	42S, 141E	Irving et al. (1974), Peck and Valley (2000)
7. Whitestone anorthosite and diorite	9/44	Hemo-Il	<980[c]	−60, 288	18S, 149E	Ueno et al. (1975), Dallmeyer and Sutter (1980)

S/s, number of sites/number of samples in study.
[a] Cooling age calculation.
[b] Regional K–Ar age.
[c] Site $^{40}Ar/^{39}Ar$ age on hornblende and hemo-ilmenite composition.
[d] See text for discussion.

suggest an apparent polar wander path that is moving in a counter-clockwise direction, and has reached its most southerly position around 980 Ma. By 960 Ma the path is moving northward to reach poles of equatorial position in the late Neoproterozoic. As can be seen in Fig. 13, the poles plotted here fall to the west of generalized paths for Laurentia, represented by one clockwise path (data from Hyodo and Dunlop (1993) and Alvarez and Dunlop (1998) as plotted by Weil et al. (2004)), and a counter-clockwise path (Weil et al., 2006). Both studies use much of the same data, but differ on interpretation of magnetization ages and weighted-importance of certain studies. The eastern limb of each loop is the most poorly defined in terms of number of poles; our data suggest that the loop may be much tighter than portrayed.

It is intriguing to compare the polar wander path from Laurentia at this time to those of other cratons, especially Baltica, which is postulated to be close to Laurentia in most reconstructions (Li et al., 2008). It was suggested early on that poles from the Sveconorwegian orogeny in Baltica, also dated around 1.0 Ga, plotted in a loop showing counter-clockwise motion (Pesonen and Neuvonen, 1981). Subsequent work has also supported a clockwise loop (Pisarevsky et al., 2003; Piper, 2009), and also the interpretation that either a clockwise or counter-clockwise loop can fit the current data and age constraints (Pisarevsky and Bylund, 2006). It is clear that both regions would benefit from additional detailed paleomagnetic studies with well-dated magnetization histories.

7. Conclusions

(1) Three distinct rock units from the Adirondack Highlands, microcline gneisses of the western Highlands, metamorphosed anorthosites and associated rocks from the central and eastern Highlands, and unmetamorphosed fayalite granite, have stable remanent magnetizations. Normal and reversed directions are represented in each group. Mean magnetic directions for the gneisses are $I = -62.8°$ and $D = 289.2°$, for the anorthosites, $I = -67.3°$. $D = 283.9°$, while the granites have distinct directions of $I = -75.8°$ and $D = 297.0°$.

(2) Studies of the magnetic minerals carrying the remanence shows different oxides are dominant in the three units. The microcline gneisses are dominated by ilmeno-hematite; of GMS sites numerous ones contain only ilmeno-hematite, while other sites have co-existing multi-domain magnetite. Hemo-ilmenite is rare in the anorthosites and the oxide mineralogy is dominated by coarse multi-domain magnetite, with minor pseudo-single domain magnetite. The Wanakena Granite has rare hemo-ilmenite, and discrete magnetite present as multi-domain and pseudo-single domain grains.

(3) Pole determinations for these three units yield positions in low to mid southern latitudes and variable longitudes, with the gneiss pole at 18.4°S, 151.1°E, the anorthosite and associated metamorphic rocks pole at 25.1°S, 149.0°E, and the granite pole at 28.4°S, 132.7°E.

(4) Using a maximum acquisition temperatures of ∼520 °C for the CRM in the ilmeno-hematite, and ∼570 °C for the TRM for the magnetite, and published cooling curves for different parts of the Adirondack Highlands yields ages of remanence of ∼990 Ma for the granite, ∼970 Ma for the anorthosite, and ∼960 Ma for the microcline gneiss. Using minimum temperature of 420 °C for the GMS unit would date the pole at ∼940. This data supports counterclockwise motion on the Grenville loop during the waning stages of the Ottawan orogeny.

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