Genesis Of The Central Tennessee Mississippi Valley-type Ore Deposits And Host Rock Dolomitization From Paleomagnetism

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Genesis of the Central Tennessee Mississippi Valley-Type Ore Deposits and Host Rock Dolomitization from Paleomagnetism

by
Michael T. Lewchuk

Department of Earth Sciences

Submitted in partial fulfilment of the requirements for the degree of Doctor of Philosophy

Faculty of Graduate Studies
The University of Western Ontario
London, Ontario
November, 1995

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Abstract

Paleomagnetic analysis was conducted on over 1100 specimens from 91 sites collected from a transect of Paleozoic carbonates in Illinois, Kentucky and Tennessee and from mineralization in the central Tennessee Mississippi Valley-type (MVT) zinc-lead ore district. Three characteristic remanence components (ChRMs), termed A, B and C, were observed. The oldest, A, was found in 24 sites of early Paleozoic, fine-grained dolostones from Tennessee and Kentucky. Its mean pole, after reversal to its antipodal position, of 126.9°E, 46.1°N, (δp=1.5°, δm=3.0°) is concordant to the “Kiaman” reversed superchron portion of the Permian apparent polar wander (APW) path for North America. The A ChRM is probably related to the pervasive regional dolomitization that completely surrounds the central Tennessee MVT ore district. The B ChRM, found in 14 sites of MVT ore-stage mineralization, is also dominantly reversed, and gives a pole position of 111.9°E, 50.5°N, (δp=1.7°, δm=3.5°). It falls on the Permian-Triassic boundary of the North American APW path and is both statistically different from, and younger than, A. Therefore, B is primary and dates the genesis of the mineralization in the central Tennessee ore district. The C ChRM was found at 19 sites in the northwestern part of the transect. Its pole position of 102.7°W, 80.8°N, (δp=5.1°, δm=6.4°) falls on the present Earth’s magnetic pole position and hence it is probably less than 10,000 years old. Successfully defining these
three ChRMs, both geographically and temporally, shows that paleomagnetism can be used to date and map fluid-flow events in sedimentary basins.

Bingham statistics were applied to the data for the A and B ChRM directions to examine the process of MVT mineralization. The azimuth and amount of bias of the A and B ChRMs show that host rock remagnetization (and probably pervasive dolomitization) predated mineralization by about 20 Ma and that the ore deposit formed in about 10 Ma. Bingham analysis for five other North American MVT ore districts also shows that the remagnetization of the host rocks was a precursor to the mineralization and that mineralization required ~25 Ma on average.
Acknowledgements

There are many people that need to be acknowledged for their help and support, both professionally and personally, during this project.

On the scientific side first and foremost is my supervisor David Symons who has provided me with as much encouragement and support as anyone could possibly ask for in a thesis supervisor. I would also like to thank both David Leach of the U.S.G.S. and Don Sangster of the G.S.C. for helping me understand the non-paleomagnetic aspects of Mississippi Valley-type ore deposits. Finally, I would like to thank all of my friends and colleagues for their stimulating scientific conversations both within and outside the field of Earth Sciences.

On the personal side I would like to thank my parents for continuously pushing me along the academic path ever since I was a young child.

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Chapter 1

Introduction

1.1 History of the Project

The main goal of this thesis is to use paleomagnetic methods to attempt to date ore deposits that do not have well constrained ages. The paleomagnetic laboratory at the University of Windsor has been active in this research area since about 1987 and has successfully dated several North American Mississippi Valley-Type (MVT) ore deposits: northern Arkansas - Tri-State (Pan et al., 1990); central Missouri (Symons and Sangster, 1991); Polaris (Symons and Sangster, 1992); Pine Point (Symons et al., 1993); Gay's River (Pan et al., 1993), and Newfoundland Zinc (Pan and Symons, 1993). The initial intent of this project was to date paleomagnetically an additional MVT ore deposit and then to try to extend the method beyond MVT ore deposits into other types of epithermal ores.

The first step was to collect samples from the Elmwood-Gordonsville MVT orebody and its Paleozoic host carbonates in the central Tennessee (CT) district during the summer of 1991. This was the only major MVT ore district
in North America not previously analyzed paleomagnetically. The collection
(26 sites, ~300 specimens) was prepared and initial measurements were
made in the winter of 1991.

During the summer of 1992 two collections were obtained from two
major mining districts for the purposes of extending the paleomagnetic
method beyond MVT ores. The first collection was of a Carlin-type gold
deposit in northern Nevada (37 sites, ~450 specimens) and the second was
from silver-bearing veins of the Coeur d'Alene district of Idaho (34 sites, ~400
specimens). These collections were prepared in the fall of 1992 and measured
during that winter along with the remaining samples from Tennessee.

Unfortunately the samples from the gold deposit in Nevada proved
to be both weakly magnetized and unstable so the collection was abandoned.
In addition there were problems with the Coeur d'Alene samples as they
contained large amounts of siderite (FeCO₃) which has a high susceptibility
but very low remanence. This caused problems with the cryogenic
magnetometer because the siderite responded to the trapped field within the
measuring coil. After numerous attempts to remove or control this effect by
modifying the measuring process, this collection was also abandoned. Details
about these two studies can be found in Appendices A and B.

The samples from the CT district proved to have much more stable
characteristic remanences. Coherent data were obtained, resulting in a pole
for the ores which fell on the Permian-Triassic boundary of the North
American apparent polar wander (APW) path. Perhaps the most important observation from this study was that the magnetization age of the Ordovician host rocks was statistically distinct but only slightly older than that of the ores when the data were analyzed using the more rigorous Bingham (1974) statistics rather than the traditional Fisher (1953) statistics. This observation differed from the conclusions of previous studies in that they generally found the same direction and age for both ores and host rocks in any given district.

Following up on the CT result, the site mean remanence directions from previous studies were re-analyzed using Bingham statistics rather than the customary Fisher statistics. By using Bingham statistics it was found that, while the ore and host rock directions were not always distinct, there was a systematic bias between the two groups with the mean pole for the host rocks being consistently older than the corresponding pole for the ores. In addition the site mean directions from the ores were streaked out along a preferential direction which could be defined and quantified using Bingham statistics. The amount of elongation was used to estimate the length of time required for the mineralizing process. The results of this study were completed in 1994 and published in the journal *Geology* in February of 1995. A copy of the article is provided in Appendix C.

A concurrent paleomagnetic study (Symons, 1994) to the CT project yielded a Jurassic direction from fluorite (MVT?) vein deposits and
surrounding host rocks in the Illinois-Kentucky (IK) fluor spar district. The IK district is only 100 to 150 kilometres away from the CT district and both are hosted by Paleozoic carbonates. It had been previously thought that these two districts were cogenetic and therefore coeval. The discrepancy between the results of these two studies, the failure of the Nevada gold and Coeur d'Alene silver projects, and the recognition that MVT ores and their host rocks retained different magnetizations prompted a detailed study of limestones and dolomites on a transect between the CT and IK districts. In the summer of 1993 samples were obtained from 16 quarries (48 sites, 571 specimens) and from an additional 13 sites from within the Elmwood-Gordonsville mine. The sites from the mine were collected to better define the shape distribution of the paleomagnetic data for the ore population. These samples were prepared and measured during the winter of 1993/94. Three characteristic remanent magnetizations (ChRM s) were recognized, termed, A, B and C, in order from the oldest to the youngest. The A and C ChRM s were isolated from the quarry samples and the B ChRM was isolated in samples from the ores and ore stage material. The results of the paleomagnetic analysis of the quarry sites are presented in Chapter 2 and the paleomagnetic results of the central Tennessee ore project are given in Chapter 3.

Finally two attempts were made to discern the hysteresis properties of the MVT ore minerals. Previous researchers (e.g. Jackson et al., 1992) had
noticed that the shape of the hysteresis curves were distinct for remagnetized carbonates when compared to carbonates which had retained their original magnetizations. Hysteresis measurements were made at the Parc St. Maur laboratory of the Institut de Physique du Globe de Paris in Paris, France, the Institute for Rock Magnetism in Minneapolis, Minnesota, as well as at the University of Toronto using the most sensitive rock magnetometers available. While a few the host rock carbonates tended to follow the previously observed behavior patterns, the majority of the carbonates and all of the ore samples were simply too weakly magnetized to yield any useful data so the experiment had to be abandoned. The results of this experiment are described in Appendix D.

1.2 The Characteristics of MVT Ore Deposits

MVT ore deposits are epigenetic base metal sulphide ore deposits which are mined for lead, zinc, barite and occasionally fluorite. Their characteristics have been summarized by Anderson and Macqueen (1982) and Sangster (1986). While the type of deposits is named after large deposits in the Mississippi Valley basin, they occur worldwide. They tend to be found in carbonate rocks along the margins of large sedimentary basins, often at basement highs. They are formed by precipitation from hot (100 to 150°C) saline formational brines which were expelled from the basin during or after its formation. They are commonly associated with oil and gas, leading to the
hypothesis that the petroleum is the source of the sulphur while the
formational brine is the source of the base metals (Anderson and Garven,
1987). This is an important point because it is virtually impossible to
transport sulphur and base metals together in solution, given the
temperature and pH restrictions of an aqueous brine passing through
carbonates. Researchers have hypothesized that either two separate
transportation pathways were used or the base metals and sulphur were
transported at different times through the same pathway (Banner et al,
1988; Fontboté and Gorzawski, 1990). However, independent evidence for
either of these two hypotheses has not been offered.

Potential driving mechanisms for the movement of the brines can be
divided into two broad categories. The first mechanism is that of fluid
migration by overpressuring due to compaction of sediments such as that
proposed by Jackson and Beales (1967) and later modified by Cathles and
Smith (1983). Under this model the fluid expulsion is a normal consequence
of basin evolution. As the basin is filled in, the weight of the overlying
sediments causes compaction of the underlying sedimentary rocks, reducing
their pore volume to cause the updip expulsion of brines from depth into the
adjacent carbonate platforms. The age of the ore deposits should be
coincident with the age of the infilling of the basin according to this model.
The second, and more popular mechanism, is that of gravity-driven fluid
migration whereby flow is generated by differences in hydraulic head across
the basin as a result of regional deformation. The most widely-quoted conceptual model is that of Garven and Freeze (1984a, b) (figure 1.1). They propose that overthrusting on one side of the basin, as a result of collisional orogenesis, causes sudden and dramatic changes to the hydraulic regime of the basin which initially drives the brine deeper below the thrust belt where it is capable of leaching metals from basal units. Eventually the hydraulic head causes the expulsion of the brines updip along permeable strata into the carbonate platform at the margin of the basin. Under this model the age of the ore deposits should be coincident with nearby orogenesis.

Extensive studies of the physical and chemical conditions during the formation of MVT mineralization have been made (Sverjensky, 1981; Cathles and Smith, 1983; Sangster, 1983; Leach and Rowan, 1986); however, obtaining a reliable age estimate on MVT ores by traditional methods has proven difficult for two reasons. First, the mineralogy of these ore deposits is simple and wall rock alteration is rare which makes it difficult to find minerals suitable for radiometric dating (Sangster, 1986). When radioactive isotopes are found in MVT minerals, they usually occur in such low concentrations (ppb range) that measurement imprecision leads to large errors or contamination by exotic materials creates problems (Brannon et al, 1995). Second, MVT deposits usually occur in tectonically stable settings so that cross-cutting post-ore geologic structures are rarely found which would enable the age of a deposit to be bracketed.
Schematic cross-section modified from Sverjensky and Garven (1992) showing the mechanism of the gravity-driven fluid migration model. Arrows indicate the sense of fluid migration away from the thrust belt into the foreland bulge on the other side of the basin.
Paleomagnetic dating of MVT deposits has been tried previously with mixed results. The first attempts were by Beales et al. (1974) and Wu and Beales (1981) but they were unsuccessful because neither the equipment nor the definition of the Paleozoic APW path were sufficient at the time. Wisniowiecki et al. (1983) obtained a Permian paleomagnetic pole for the Pb-Zn ore deposits of southeast Missouri. They argued that the magnetization was older than the mineralization based on geochemical considerations. Bachtadse et al. (1987) found a Permian pole for the structurally-complex zinc deposits of the east Tennessee (ET) zinc district. In contrast they argued that the magnetization postdated the mineralization based on the results of a paleomagnetic fold test.

More recently researchers in the paleomagnetic laboratory at the University of Windsor have measured several MVT ore deposits throughout North America, and it has been argued for each district that the age of mineralization is coeval to the age of magnetization. The results of the paleomagnetic studies at the University of Windsor paleomagnetic laboratory (table 1.1) fit the conceptual model of Garven and Freeze (1984 a, b) because all of the ore deposits yield magnetization ages which are coeval to the age of major events in nearby orogenic belts. This is true despite the wide geographic distribution of the studies (figure 1.2) and the varied host rock ages. The only exception to this trend may be the fluor spar vein deposits of the IK fluor spar district. Their paleomagnetic age does not coincide with any
<table>
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<tr>
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<th>Paleomagnetic age</th>
<th>Orogen</th>
<th>Orogeny</th>
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</thead>
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<tr>
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<td>Late Devonian</td>
<td>Innuitian</td>
<td>Ellesmerian</td>
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<tr>
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<td>Middle Devonian</td>
<td>Late Cretaceous/ Paleocene</td>
<td>Cordilleran</td>
<td>Laramide</td>
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<td>Alleghenian</td>
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<td>late Penn./Early Permian</td>
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<td>Ouachita</td>
</tr>
<tr>
<td>Central Missouri</td>
<td>Early Ordovician</td>
<td>Late Permian</td>
<td>Ouachitan</td>
<td>Ouachita</td>
</tr>
<tr>
<td>Northern Arkansas</td>
<td>Ordovician &amp; Mississippian</td>
<td>Permian</td>
<td>Ouachitan</td>
<td>Ouachita</td>
</tr>
</tbody>
</table>

**Table 1.1**

Host rock and paleomagnetic ages, along with their associated orogen and orogeny, for the North American MVT districts that have been studied paleomagnetically. In all cases the paleomagnetic age agrees well with the timing of a nearby orogenic event. Pennsylvanian (Penn.) For references to the individual studies see sections 1.1 and 1.2.
Locations of the North American MVT ore deposits which have been paleomagnetically dated. The shaded areas are the approximate boundaries of the orogenic belts thought responsible for the brine migration and MVT mineralization according to the hypothesis of Garven and Freeze (1984a, b).
nearby orogenic activity but it is also quite likely that they are not true MVT deposits.

One concern of this method of dating MVT deposits has been whether or not the magnetization is truly coeval with the mineralization. This thesis will attempt to put that concern to rest through the detailed analysis of both the ores and their host rocks in the CT district in addition to dating and estimating the duration of the genesis of the ore. This thesis will also attempt to define extent of the fluid-flow events responsible for the formation of the ore in this district.

1.3 Equipment and Laboratory Methods

The paleomagnetic analysis of MVT ore deposits requires extremely sensitive equipment and detailed analytical procedures because these rocks are usually extremely weakly magnetized. For this reason all paleomagnetic analyses were carried out in the University of Windsor paleomagnetic laboratory which has one of the most sensitive and efficient magnetometers in the world.

All measurements were conducted within a magnetically-shielded room which consists of three layers of transformer steel on all sides including the roof and floor. Each layer is separated from the next by about 10 cm. The Earth's magnetic field is preferentially channeled parallel to the plane of the walls thereby reducing the magnetic field within the room to less than 0.2%
of the ambient Earth's magnetic field. This shielding also protects the sensors from electromagnetic interference such as television and radio carrier waves, cellular phone transmissions and CB radios.

The specimens were measured on a two-axis Canadian Thin Films (CTF) cryogenic magnetometer which is fully automated and interfaced to a computer that controls both the movement of the specimen holder as well as the acquisition of data. The CTF magnetometer, when operated within the confines of the shielded room, is one of the most sensitive machines in the world. Unfortunately, even by MVT standards, the samples from the CT district were very weakly magnetized and several changes and upgrades to the equipment were made over the course of this thesis. The specimens were initially measured using an Apple 2+ computer and software that had not been substantially upgraded since the early 1980's. Many of the specimens in the collection had initial natural remanence magnetization (NRM) intensities that were very near the noise level of the equipment (~2 x 10^{-5} A/m for a standard 2.5 cm diameter by 2.2 cm height right-cylindrical drill core specimen). This indicated that there would be problems obtaining enough reliable data for a useful result. Therefore software was written for an IBM 486 compatible computer to take advantage of the advances in computer technology over the last several years. The main improvement was in the way the program was able to store and subtract both zero and sample holder magnetizations, thereby lowering the background level. Also the new
program, using the much faster speed of the new computer, was able to measure each axis up to 100 times as opposed to the original program which could only take one reading on each axis. This resulted in much more reliable data when a specimen’s magnetization was close to the sensitivity threshold of the magnetometer. It also resulted in a significant decrease in the time required to complete a measurement cycle. Also, at the urging of a visiting professor, Dr. Kazuto Kodama, the magnetometer electronics were put on a separate power source, and a power stabilizer was installed. The net result of these improvements to the equipment was about a five fold increase in sensitivity ($\sim 4 \times 10^{-6}$ A/m for a standard specimen) to a machine that was already one of the best available. Signal stacking, frequent tuning of the sensors and meticulous cleaning of the sample holder allowed for the reliable detection of specimen NRM$s$ as low as $7 \times 10^{-7}$ A/m on occasion. Despite these changes a large portion of the collection from the ore deposit had to be abandoned because it was still too weak to obtain reliable data.

NRM demagnetization was conducted using a Sapphire Instruments SI-4 alternating field (AF) demagnetizer with a maximum field of 160 mT or a Magnetic Measurements MMTD-1 thermal demagnetizer with a maximum temperature of over 700°C. Saturation isothermal remanence magnetizations (SIRM’s) were imparted using a Sapphire Instruments SI-6 pulse magnetizer. All of this equipment is housed within the shielded room so that
it was not necessary to remove the rocks from the shielded environment at any time during the measuring process.

1.4 Statistical Methods

Dispersion within paleomagnetic data can be caused by both random and nonrandom sources. Random dispersion is a function of sampling and measurement imprecision plus short term (<10^-2 Ma) secular variation of the Earth's magnetic field, and it can be considered to be the true error. Random dispersion should be the same in all directions around the mean. Nonrandom dispersion can be caused by several factors such as structural distortion, rock fabric and APW. Nonrandom dispersion will be reflected as a bias in the shape distribution of the paleomagnetic data set. Fisher (1953) statistics assume that dispersion in a population of paleomagnetic data is caused solely by random error so it only accounts for the magnitude of the variation about the mean, not the degree of bias or its direction.

Bingham (1974) statistics assume that both random and nonrandom variation can be present in a paleomagnetic population and can, therefore, describe both the amount of dispersion and its direction of bias. Traditionally Fisher statistics have been used by paleomagnetists despite knowing that Bingham statistics provide a better description for many paleomagnetic populations. The reasons for this are that Fisher statistics can be easily
computed using simple computer programs, that their interpretation is straightforward, and that the calculated confidence circles give conservative error estimates when compared to Bingham statistics. Thus Bingham statistics have been avoided except where necessary. This thesis will present a case where they yield both a better interpretation of the data and also provide additional information that Fisher statistics do not recognize.

Fisher statistics assume that the unit vectors are randomly distributed about their mean with a Gaussian distribution away from the mean in all directions. When analyzed in site mean ChRM coordinates, Fisher statistics will generate a confidence cone about the mean which is circular to give a circle of 95% confidence (radius=\(\alpha_{95}\)) (figure 1.3). When the circle of confidence is converted into polar coordinates it forms an ellipse according to the dipole formula (Tarling, 1983). The orientation of the axes of the ellipse, \(\delta p\) and \(\delta m\), are controlled by the relationship between the location of the sampling site and its pole such that \(\delta p\) is along the plane of the meridian between the site and the pole and \(\delta m\) perpendicular to the meridian. Therefore the orientation of the Fisher ellipse, in polar coordinates, is in no way related to the shape distribution of the magnetization population. The generation of an ellipse in polar coordinates can be avoided by converting individual site mean ChRMs into virtual geomagnetic poles (VGP's) and then analyzing the VGP's using Fisher statistics. The confidence in polar coordinates will then be given by a circular cone of 95% confidence.
Hypothetical site mean distribution (a) which is skewed along a trend of apparent polar wander and the resulting mean (b) circumscribed by both its circle of 95% confidence after Fisher (1953) and oval of 95% confidence after Bingham (1974). Notice that $\alpha_{1.2}$ of the Bingham ellipse is longer than $\alpha_{1.3}$ and the trend of $\alpha_{1.2}$ is parallel to the trend of apparent polar wander in (a).
(A_{92}). Both methods are considered acceptable uses of Fisher statistics although the former is more commonly used.

Bingham statistics recognize that the unit vectors of a population can be described with respect to their mean by both the magnitude of deviation as well as the azimuth from the mean to the individual data point. Therefore, the confidence ellipse defined by Bingham statistics will have a biased or elliptical distribution relative to its mean that is directly related to the shape distribution of the population about the mean. The 95% confidence interval generated by Bingham statistics is an ellipse with a semi-major axis of α_{1,2} that also defines the bias direction, and a semi-minor axis of α_{1,3} that is perpendicular to α_{1,2} (figure 1.3). If the population is randomly distributed (i.e. a true Fisherian distribution), then α_{1,2} and α_{1,3} will be equal. However, if there is bias in the distribution, then the amount will be reflected by the difference in magnitude between α_{1,2} and α_{1,3}. The direction of the bias will then be represented by the azimuth, expressed in degrees from north, of α_{1,2} with the azimuth being defined as a line tangential to the unit sphere at the coordinates of the mean and parallel to the direction of bias in the distribution. When remanence directions are distributed along a linear trend, their skew will be reflected by the Bingham but not by the Fisher statistics. Elongation along a great-circle path between the present Earth's magnetic field (PEMF) pole and the ChRM direction could be evidence of only partial removal of a viscous overprint by incomplete AF or thermal demagnetization.
Elongation along any other great circle could indicate a sampling or measurement bias, a structural bias from folding or faulting, or else it could be evidence for a prolonged acquisition time for the component so that it records a portion of the APW path.

MVT deposits normally occur in flat-lying undeformed sediments, thus precluding the possibility of a metamorphic or tectonic origin for the bias. In addition the paleomagnetic collections of MVT deposits previously analyzed in the Windsor lab originated from oriented hand samples because gasoline powered drills are not allowed underground for safety reasons. This method of collecting tends to randomize the sample declinations and inclinations, which prevents the artificial introduction of sampling or measurement bias into the data when compared to field drilling which results in vertically-biased sample orientations. Therefore, the most plausible mechanism for biased distributions in most MVT deposits is the recording of APW. This possibility can be tested by comparing the azimuth of elongation in a given population to the general trend of the of APW path for the time period indicated by the population mean. If the bias in the population is the product of APW during the acquisition of the ChRM, then the trend of the bias will approximate the trend of the APW path.

If the source of the bias is the recording of APW then a simple estimate of the remanence acquisition time can be made using Bingham statistics. Both the $\alpha_{1,2}$ and $\alpha_{1,3}$ axes account for random dispersion but only
the major $\alpha_{1.2}$ axis also includes dispersion from APW expressed in degrees. If the rate of APW can be expressed in degrees per unit of time, then the duration of remanence acquisition can be obtained. Following Lewchuk and Symons (1995) the duration of magnetization acquisition can be calculated using the formula:

$$2(\alpha_{1.2} - \alpha_{1.3}), \times \text{rate of APW} = \text{duration}$$

where $\alpha_{1.2}$ and $\alpha_{1.3}$ are expressed in degrees and the rate of APW is expressed in Ma/degree. For the Paleozoic the best way to estimate the rate of APW is to use the reference poles of Van der Voo (1993). Each reference pole has both a direction and an age. By measuring the angular deviation between reference poles and their difference in age, a rate of APW in millions of years per degree of arc (Ma/$^\circ$) is obtained. The average rate of APW during the time when these ore deposits formed, i.e. between 340 and 235 Ma, is $2.7 \pm 0.7$ Ma/$^\circ$. This method has already been applied by Lewchuk and Symons (1995)(appendix C) to several other MVT ore deposits. The results for three of those deposits, from three different basin-orogen pairs, are shown in figure 1.4. Notice that in each of these cases, the host rock pole (diamond) falls on an older portion of the APW path than its corresponding ore pole (circle). Also notice that in all three cases the confidence ellipse about the ore pole is elongated along a trend that tends to align with the trend of the APW path for the time indicated by its mean, indicating that the bias is the product of APW. Finally, according to the method of Lewchuk and Symons (1995), an
Mean pole positions of three North American MVT ore districts, adapted from Lewchuk and Symons (1995; also appendix C), plotted on the Paleozoic apparent polar wander path for North America (Van der Voo, 1990). Circles are the ore mean poles while diamonds are the host rock mean poles. The star represents the published combined pole where the original results differ from those presented here. Su, D1, Dm, Du, Cl, Cu, Pl, Pu and Trl refer to the Late Silurian through Early Triassic epochs.
estimate for the duration of acquisition can be made for each study. While the method is probably not rigorous enough to differentiate between one deposit and another, generalizations about the average time that it takes to form a typical MVT deposit can be estimated. Lewchuk and Symons (1995) have argued that a typical MVT ore deposit took about 25 Ma to form.

This method for estimating the acquisition time will be tested in the CT ore district through the use of a detailed study of both the ores and host rocks. Previous studies had only been able to provide rough estimates of the duration of the mineralizing process because fewer samples of both MVT mineralization and host rocks were taken when compared to this project.
Chapter 2

Paleomagnetism of Paleozoic carbonates in the Illinois-Kentucky-Tennessee area: the recording of fluid flow events

2.1 Introduction

Numerous paleomagnetic studies on early Paleozoic sedimentary rocks in the midcontinent of the United States have yielded secondary, late Paleozoic directions corresponding to the “Kiaman” reversed superchron (e.g. McCabe and Elmore, 1989). Recent paleomagnetic investigations on MVT ore deposits in the midcontinent have also yielded late Paleozoic directions (Pan et al., 1990; Symons and Sangster, 1991). These magnetizations are primary in the ore minerals, thus providing the first reliable age dates for the MVT deposits. The ages obtained support the theory that MVT ores were formed from the mass migration of basinal brines (Leach, 1979; Ohle, 1980; Anderson and Macqueen, 1982; Oliver, 1986, 1992) into the midcontinent during the late Paleozoic when fluids were expelled from nearby orogenic
uplifts (Garven and Freeze, 1984a, b). These, or related fluids, are also thought responsible for the remagnetization of the host rocks to give magnetization ages that are similar to the resident ores in virtually every study despite very different depositional ages for the host rocks. This has led to the conclusion in previous studies that the ore and host rock magnetic directions are the same. However, combined analysis of the paleomagnetic data from several North American MVT districts using Bingham (1974) statistics, has shown that the host rock magnetizations are slightly older than the magnetizations of their mineralization (Lewchuk and Symons, 1995)(appendix C).

One ore district that has a paleomagnetic age which does not coincide with the orogenic uplift model is the Illinois-Kentucky fluorspar (IK) district. It gives a Jurassic paleomagnetic direction for both ores and nearby host rocks (Symons, 1994). This ore district is only ~150 km from the central Tennessee zinc (CT) district as well as from the Old Lead belt and Viburnum trend (New Lead belt) districts of southeast Missouri (figure 2.1), all of which have late Paleozoic paleomagnetic signatures. Hayes and Anderson (1992) suggested that the IK district is coeval with the CT district based on similarities found in fluorite banding which, if true, would show that the IK district is the product of MVT ore forming fluids and thus indicate a problem with the paleomagnetic age for the IK district. Others have argued that the IK district is not a true MVT district (Leach and Sangster, 1993; Symons,
Figure 2.1

Regional map of the study area, showing the major geological structures and their spatial relationship to MVT ore districts. District abbreviations: TS=tri-state; CM=central Missouri; NA=northern Arkansas; SeM=Southeast Missouri (Old and New Lead belts); IK=Illinois-Kentucky; CT=central Tennessee; CK=central Kentucky; and ET=east Tennessee. Heavy rectangle is approximate area of figure 2.2.
1994), but rather that it is a product of rift-related tectonics and may be genetically tied to the development of the Reelfoot rift. If the uplift model for MVT mineralization is correct and the paleomagnetic ages are valid, then the IK district is not a true MVT district. Also the ore and host rock magnetizations in the IK district require an additional mineralizing event of unexplored extent in the midcontinent, the hydrothermal Rosiclare event (Symons, 1994).

This research was conducted in an effort to define the boundaries of the Jurassic IK district event as well as to provide additional data to better define the host rock magnetizations of the CT district, a more traditional MVT setting.

2.2 Geology

The geology of the study area has been described in detail by Kyle (1976) and Schwalb (1982). It consists primarily of subhorizontal Paleozoic carbonates overlying the Precambrian basement. Minor post-Paleozoic sedimentation is largely restricted to the Mississippi embayment and the Illinois basin. There are several major structural controls on Paleozoic sedimentation. Structural highs of the Nashville (Tennessee) and Jessamine (Indiana, Kentucky, Ohio) domes form the southern extension of the Cincinnati arch, a broad anticlinal structure (figure 2.1). The domes are separated by a sag known as the Cumberland saddle. The Illinois basin,
Reelfoot rift and Rough Creek-St. Genevieve grabens were the main
Paleozoic structural lows (figure 2.2). The Reelfoot rift, which started
developing in the late Precambrian or earliest Cambrian, extends from
southern Illinois into Arkansas. It consists of a graben bounded by the
Nashville dome on the east and the Ozark dome on the west. The Pascola
arch, a nearly circular uplift of uncertain tectonic origin that formed in the
latest Paleozoic or Mesozoic, lies between the Ozark and Nashville domes
and it forms the southern flank of the Illinois basin (figure 2.1). North of the
Pascola arch and, also of Mesozoic age, are the Rough Creek graben and the
St. Genevieve fault zone. These structures are similar to the Reelfoot rift but
smaller. The Rough Creek graben extends eastward from the IK mining
district across the Cumberland saddle into Kentucky whereas the St.
Genevieve fault zone extends westward between the Ozark dome and the
Illinois basin (figure 2.2). Uplift of the Ozark dome was re-initiated in the
Mesozoic, Tertiary and Recent time periods (Schwalb, 1982).

The basal sedimentary unit is normally the Lamotte/Mt. Simon
sandstone of the Potsdam megagroup. It rests unconformably on the
Precambrian basement everywhere except in the middle of the Reelfoot rift
where some pre-Potsdam sedimentary rocks are found at depth. The
uppermost rock units in the southern part of the study area are
Cambro-Ordovician carbonates, principally the Knox megagroup which
consists of a thick sequence of carbonates with shales and sandstones
Figure 2.2

Locations of the sixteen quarries sampled in this study. District abbreviations as in figure 2.1. Quarries sampled as follows: 1=Dravo Basic Materials, Hardin County, IL; 2= Hardin County Materials, Hardin County, IL; 3=Jan Stone Enterprises, Eddyville, IL; 4= Columbia Quarry Company, Columbia, IL; 5=Southern Illinois Stone Company, Buncomb, IL; 6=Anna Quarries Inc., Anna, IL; 7=Vulcan Materials, Gilbertsville, KY; 8=Kentucky Stone Company, Princeton, KY; 9=Kentucky Stone Company, Marion, KY; 10=Dravo Basic Minerals, Birdsville, KY; 11=Kentucky Materials, Canton, KY; Vulcan Materials, Lebanon, TN; 13=Vulcan Materials, Readyville, TN; 14=Vulcan Materials, Hermitage, TN; 15=Vulcan Materials, Clarksville, TN; 16= Porter Brown Limestone Company, Pleasant View, TN.
marking its base. This unit rests conformably on the Potsdam. The Knox generally thickens away from the Nashville dome, a paleohigh at the time of deposition, to exceed 2000 m in the Reelfoot rift. The carbonates have been extensively dolomitized by both meteoric and basinal fluids (Montañez, 1995). The Knox was exposed to erosion and extensively karstified in the Ordovician although karstification may have continued until the end of the Paleozoic (Schwalb, 1982). Knox carbonates host the MVT ores of both the CT and east Tennessee zinc (ET) districts. While the Knox megagroup is the uppermost unit at the crest of the Nashville dome and Pascola arch, it is unconformably overlain by the Mississippian Warsaw and St. Louis formations elsewhere in Tennessee, western Kentucky and southern Illinois. Dominantly carbonate sedimentation continued through to the end of the Paleozoic in the northern part of the study area, particularly as subsidence continued in the Illinois basin, forming a cyclical carbonate-sandstone-shale sequence of over 1000 m in thickness in southern Illinois and thinning towards the late Paleozoic structural highs to the south. The Pennsylvanian St. Genevieve formation is a major source for construction materials that are exploited by quarries in both Kentucky and Illinois. Extensive coal units were laid down in Pennsylvanian time also. Although their coal rank indicates that they were deeply buried, their rank may, in fact, be the product of hot, upwelling basinal brines which may be responsible also for
dolomitization, elevated conodont alteration indices (Sangster et al., 1994) and/or MVT mineralization (Bethke and Marshak, 1990).

Uplift of the Appalachian and Ouachita mountains associated with the Alleghenian/Ouachitan orogeny occurred in the Pennsylvanian and Early Permian, coinciding with the convergence of Laurentia, South America and Africa. This uplift resulted in the expulsion of formational fluids into the midcontinent from the Arkoma basin to the southwest and Black Warrior basin to the southeast (Oliver, 1986, 1992). The Reelfoot rift was likely a boundary, keeping Black Warrior basinal fluids to its east and Arkoma basinal fluids to its west. This mass fluid migration has been held responsible for dolomitization, ore precipitation and elevated coal rank throughout the midcontinent. Fission track results show that the rock temperatures exceeded the 60 to 150°C annealing temperature of apatite (Arne, 1992), but have never been much warmer than about 150-200°C from fluid inclusion temperatures and conodont alteration indices (Sangster et al., 1994, and references therein).

Although numerous small dikes and diatremes have been found in the district, particularly around the Permian Hicks dome near the IK district, there has been no evidence of major igneous intrusive activity in the area that would be capable of regionally elevating rock temperatures to a level of 400°C which is necessary in order to thermally reset their remanent magnetizations.
2.3 Methods

All samples, except those from the road cut at site 48, were collected from quarries along a transect through the CT and IK districts (figure 2.2). Quarries were sampled to avoid oxidization from modern weathering. Three sites were taken at each of the 16 quarries except for the small sandstone quarry (quarry # 3, figure 2.2) where only two sites were obtained. Where possible, one of the three sites was from the bottom of the quarry, usually more than 75 m below surface. A second site was obtained from the mid level of the quarry and a third from near the top but well below the visible weathering profile. Four oriented hand samples were collected at each site and three cores were drilled from each sample in the laboratory to yield a total collection of 571 specimens.

The specimens were stored in the magnetically-shielded room for more than two months prior to measurement in order to reduce the amount of viscous remanence magnetization. All remanence measurements were conducted within the room using the CTF cryogenic magnetometer. Detailed AF demagnetization was conducted on the NRM of one pilot specimen from each hand sample. These pilot specimens were treated in 12 to 15 steps up to a peak field of 150 mT. A second pilot specimen from each hand sample was thermally demagnetized in 10 to 12 steps up to a maximum of 690°C. Most thermal specimens were abandoned by 475 to 500°C due to the alteration and creation of new magnetic minerals during heating. The remaining
specimens were AF demagnetized in five to seven steps to a peak field of 150 mT based on the results of the pilot specimens. Specimen ChRM directions were obtained using the method of Kirschvink (1980) and were accepted for statistical analysis if the best-fit stable direction had a mean angular deviation of <15° over three or more steps. Site mean directions were calculated using the statistical method of Fisher (1953). Overall population mean directions were described using both Fisher (1953) and Bingham (1974) statistics. The latter were preferred because they better describe the dispersions of the populations.

Two specimens from each quarry were subjected to SIRM testing by using a direct field pulse magnetizer to progressively impart an isothermal remanent magnetization in each specimen in 11 steps up to a peak field of 900 mT. The specimens were then AF demagnetized in 5 steps to 50 mT. The remanence magnetization was measured after each treatment and the data were normalized to the saturation intensity for the SIRM plots. Cisowski (1981) crossovers were also generated by plotting isothermal remanent magnetization (IRM) acquisition vs. DC field treatment and IRM intensity decay vs. AF treatment together for the same specimen. The point at which the two curves crossover can be used to help describe the magnetic mineralogy in terms of mineral type, grain size, and degree of interaction between grains. An additional specimen from each quarry was saturated in a
900 mT field and then thermally demagnetized in six steps to 600°C to see if goethite or hematite was present in the samples.

2.4 Results

Two stable ChRM directions were isolated along the traverse from central Tennessee to southern Illinois. For the most part the same ChRM direction was isolated in all three sites within an individual quarry after the removal of a viscous component below 20 mT although some sites had a higher percentage of unstable specimens without ChRM directions. A few sites showed the two components overlapping but never reaching a stable end-point ChRM direction. A south-southeast and equatorial ChRM direction or A component was found in 23 sites from eight of the 16 quarries and the road cut site. When present the A component was very stable and persistent, being observed in >90% of the specimens from these sites at demagnetization fields ranging from 20 mT to as high as 150 mT. A north and steeply-down ChRM direction or C component was found in 19 sites from the other eight quarries. When isolated, the C component was not as well defined as the A component with more specimens being rejected during statistical analysis because of high mean angular deviations. The remaining six sites exhibited streaked distributions from hybrid A-C component directions with occasional specimens from a given site retaining only the A or C component.
2.4.1 A Component

Specimens retaining the A ChRM component had a median initial NRM intensity of \( \sim 1.5 \times 10^{-4} \) A/m with median destructive fields between 40 and 50 mT (figure 2.3). The A component was isolated after removal of a steeply-downward component, presumably viscous remanence recording the PEMF. The A direction was observed in 261 of the 287 specimens measured from these 24 sites (table 2.1). Two specimens from separate blocks from quarry #7 yielded the antiparallel direction (figure 2.4a). The A component was found after both AF treatment at fields as high as 150 mT (figure 2-4b, c, d) as well as after thermal treatment to temperatures around 400°C (figure 2-4e, f, g). There was no significant NRM intensity drop in the 80 to 120°C unblocking temperature range of goethite or the 280-320°C unblocking range of pyrrhotite which eliminates them as possible carriers of the ChRM. Above 400°C the directions commonly became erratic and the intensities increased by more than an order of magnitude from the generation of magnetic minerals during further heating. Therefore it was not possible to preclude the possibility of hematite contributing to the remanence from stepwise thermal demagnetization of the NRM alone. However, the fact that AF treatment of 150 mT almost completely destroyed the NRM of the specimens argues that hematite, if present at all, is not a major contributor to the remanence.
Figure 2.3

Mean and quartile values of the A and C component NRM intensity as a function of demagnetizing field. The squares (circles) are the A (C) component and the X indicates the median destructive field.
<table>
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<tr>
<th>Site</th>
<th>Quarry</th>
<th>$N_{M} - N_{R} - N_{N}$</th>
<th>Dec.</th>
<th>Inc.</th>
<th>$\alpha_{95}$</th>
<th>$\kappa$</th>
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<td>4.4</td>
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Table 2.1

Site mean statistics from those sites retaining the A component. Quarry abbreviations as in figure 2.1; $N_{M} - N_{R} - N_{N}$ are the number of specimens measured, those with reversed components and those with normal components respectively; Dec and Inc are the declination and inclination: $\alpha_{95}$ is the cone of 95% confidence and $\kappa$ is the precision parameter (Fisher, 1953).
Figure 2.4

Orthogonal demagnetization plots of pilot specimens retaining the A component. Circles are projections in the N-E-S-W horizontal plane. Triangles are projections in the Up-N-Down-S vertical plane. AF pilots, with treatments in milliTesla (mT) are shown in a (normal polarity) and b, c, and d (reversed polarity). Thermal pilots (all reversed) with treatments in degrees Celsius are shown in e, f, and g (next page).
Figure 2.4 cont.
Figures 2.5a and 2.5b show the SIRM acquisition and decay curves of 17 representative specimens from those sites which yielded clear A component directions. Both sets of curves indicate that hematite is not present and, therefore, that single domain (SD) to pseudosingle domain (PSD) magnetite is the sole contributor to the NRM. Figure 2.5c shows the normalized intensity of nine specimens after saturation as a function of stepwise thermal demagnetization. All specimens but one were well behaved and dropped below 1% of saturation intensity by 600°C, indicating again that magnetite is the dominant remanence carrier. The absence of a distinct “knee” in the curve in the 280 to 320°C range further eliminates pyrrhotite as a carrier of the NRM. The one aberrant specimen was unstable above 300°C and was probably contaminated by the generation of magnetic minerals during heating. Cisowski (1981) crossover points are shown in figure 2.6 for the same 17 A component SIRM specimens as in figures 2.5a and 2.5b. Most specimens give a crossover at between 45 and 50% of the saturation intensity and between 38 and 52 mT. This is clearly in the range of non-interacting SD to PSD magnetite which is the most stable form, paleomagnetically, of magnetite. These four tests together with the step demagnetization results show that high-coercivity SD to PSD magnetite is the only contributor of consequence to the NRM.
Saturation isothermal remanence magnetization curves of example specimens from sites retaining the A component (a, b, and c) and the C component plus hybrids (d, e and f) (next page). Standard acquisition (a and d) and decay curves (b and e) show normalized intensities as a function of treatment field in milliTesla (mT). Thermal decay of SIRM curves (c and f) show normalized intensity as a function of temperature in degrees Celsius. The dashed lines show the expected behaviour for single domain (SD) pseudosingle domain (PSD) and multidomain (MD) magnetite as well as coarse (CH) and fine (FH) grained hematite.
Figure 2.5 cont.
Cisowski (1981) crossover points of SIRM test specimens. Circles indicate A component sites while squares indicate C component and hybrid sites. Dashed line is the theoretical upper limit for pure single domain magnetite.
2.4.2 C Component

The C ChRM component identified in this study was not as clearly isolated. Specimens retaining the C component were characterized by slightly lower median NRM intensities (~1.0 x 10^{-4} A/m) and markedly lower median destructive field values (~20 mT) than those of the A component specimens (figure 2.3). This resulted in the specimens having greater directional scatter or instability as the ChRM intensities were often close to the sensitivity threshold of ~4 x 10^{-6} A/m for the magnetometer. The C component is characterized by steeply-downward northerly directions that are similar to the PEMF direction. It was isolated in 19 sites from eight quarries although only 169 of the 224 specimens from these sites retained the C component in two or more steps. The remainder of the specimens were either unstable, too weak, or had aberrant magnetization directions. Since the C component and PEMF directions are similar, all data below the 20 mT step on AF demagnetization were rejected to ensure that only stable, higher-coercivity remanences were being analyzed. At coercivities of ~40 mT the C ChRM is as stable as A but only ~1/6th as intense (figure 2.3). Although the median destructive fields were lower for the C than A component, the C ChRM was still observed in fields as high as 150 mT (figure 2.7a-d), leading to the conclusion that this is a very stable, high-coercivity component despite its similarity in direction to a viscous PEMF overprint. As for specimens retaining the A component, thermal
Orthogonal demagnetization plots of pilot specimens retaining the C component. Projection style and symbols as in figure 2.4. AF pilots, with treatments in milliTesla (mT) are shown in a, b, c and d. Thermal pilots with treatments in degrees Celsius are shown in e, f, and g (next page).
Figure 2.7 cont.
demagnetization above 450 to 500°C caused new magnetic minerals to form that obliterated the NRM making further thermal demagnetization fruitless (figure 2.7e-g).

SIRM analyses of specimens retaining the C component yielded similar results to those from the A component specimens. Although the SIRM curves for the C component are more scattered, they still indicate that SD to PSD magnetite is the sole remanence carrier of C (figure 2.5d, e). Thermal demagnetization of SIRM (figure 2.5f) in C specimens indicate that, as in A specimens, magnetite is the sole remanence carrier. The Cisowski (1981) crossover points are slightly lower or from ~0.40 to 0.45 for C specimens (figure 2.6), as is expected from their somewhat lower median destructive fields, but still indicate that noninteracting SD to PSD magnetite is the dominant remanence carrier. In general, despite drastically different ChRM directions the A and C components yielded remarkably similar curves in the SIRM tests.

The SIRM results of this study agree with recent rock magnetic studies on carbonates from the midcontinent (e.g. Jackson, 1990; Jackson et al., 1992; Jackson et al., 1993, Suk et al, 1993, Xu et al, 1994). These studies have concluded that, while both fine-grained magnetite and coarse-grained pseudoframboidal magnetite replacing pyrite are present, it is only the fine magnetite fraction that is capable of retaining a stable remanent magnetization.
2.4.3 Hybrid Component

Several specimens from quarries in Illinois and Kentucky failed to yield clear A or C component directions although demagnetization of the NRM did reveal consistent stable ChRMs. These specimens followed a great circle track from the C component direction towards the A component direction upon demagnetization (figure 2.8), indicating the presence of residual A component magnetization. All of these specimens came from sites in quarries that had other sites which clearly retained the C component, and they possessed rock magnetic properties (SIRM, NRM intensity and median destructive field) identical to specimens retaining the C component. Since their stable end-point directions varied from specimen to specimen depending on the amount of C component overprinting, site and component means were not computed for this hybrid component.

2.5 Discussion

2.5.1 A Component

The A component was the only stable high coercivity direction that was observed in the road cut at site 48 and in seven of the 16 quarries studied (figure 2.9, open circles). It was also observed in two of three sites from quarry #10. Together these 24 sites form an elongate population (figure 2.10a, b)(table 2.1) with a mean of declination (D)= 157.2°, inclination (I)=...
Orthogonal demagnetization plots of pilot specimens with hybrid magnetizations. Projection style and symbols as in figure 2.4. AF pilots, with treatments in milliTeslas (mT) are shown in a, b, c and d. Thermal pilot with treatments in degrees Celsius is shown in d. Stereonet projection shows the A and C component means as stars along with the step demagnetization data from a, b, c and d.
Figure 2.9

Geographic distribution of quarries according to their ChRM components. Open (solid)[shaded] symbols show the quarries which retained the A (C) [C as well as the hybrid A-C] as their remanence component.
Figure 2.10

Stereonet projections of the A and C component sites (circles) and their means (stars) circumscribed by their cones of 95% confidence after Fisher. Part a is a standard projection while part b is a projection in the vertical plane to better demonstrate the elongate nature of the A component. The arrow shows the great circle path from the PEMF to the A component mean. Note that the elongation of the A component population is not within this plane.
4.4°, radius of 95% confidence (αg) = 3.0°, precision parameter (κ)=99 which corresponds to a pole position, after reversing to its antipodal position, at 126.9°E, 46.1°N (δp=1.5°, δm=3.0°)(Fisher, 1953) (figure 2.11a, b). The origin of the A component mean is straightforward. It is identical in both direction and polarity to the late Paleozoic portion of the APW path. This direction has been observed in numerous studies of Paleozoic sediments in the midcontinent (e.g. McCabe et al, 1989). Previous studies of late Paleozoic, magnetizations have yielded exclusively reversed polarities, leading to the acceptance of a “Kiaman” reversed superchron from the mid Mississippian to latest Permian (Van der Voo, 1993). Since two specimens from two separate blocks yielded antiparallel normal directions and a Permian direction was obtained, this magnetization was probably acquired at the end of the Kiaman reversed superchron when the field was predominantly but not exclusively reversed.

The A component population is elongated and thus better described using Bingham (1974) statistics. This results in a mean direction of D=157.2°, I=4.4°, α1.2=3.5°, α1.3=1.9°, and azimuth= 123.6°. The azimuth of bias or elongation of this population is important. The bias in the A population is clearly oblique to the great circle path between the PEMF and the A component mean directions (figure 2.10b), thereby excluding the incomplete removal of a viscous remanence as a possible source of bias in the population. It is unlikely that sampling bias caused this distribution because
a) Phanerozoic apparent polar wander path for North America (Van der Voo, 1990) showing the C component circumscribed by its Fisher cone of 95% confidence and its relationship to both the Jurassic component in the Illinois-Kentucky (IK) fluor spar district of Symons (1994) and the PEMF pole and b) Paleozoic portion of the apparent polar wander path from Van der Voo (1990) showing the A component mean circumscribed by its Bingham ellipse of 95% confidence. Bold letters refer to the early (e) middle (m) or late (l) epochs for Devonian (D) through Tertiary (T) epochs. Notice that the elongation is almost parallel to the trend of the path.
all the samples were collected in a single field trip using a single compass. Bias as a result of measurement is also unlikely because several other paleomagnetic collections were measured simultaneous to this one and they did not yield similar distributions. The rocks in the study area are undeformed which precludes a tectonic origin for the bias. Thus the only viable explanation is that the elongation was caused by APW during acquisition of the magnetization. Figure 2.11b shows the A component mean and trend of elongation superimposed on the Paleozoic APW path which demonstrates that the bias in A is similar to the trend of APW. Using the method described in Chapter 1 and an average rate of APW for the late Paleozoic of 2.7 Ma/°, calculated from reference poles of Van der Voo (1990), the A component data indicate that this remanence was acquired over a period about 8 to 9 Ma. This is a minimum estimate because it is likely that the presence of a host rock remagnetizing fluid over an extended period of time would continuously update the rock's magnetization. Thus the age and duration obtained by this calculation likely records only either the latter portion or the most pervasive portion of the entire remagnetizing event which may have lasted much longer.

While it is attractivc to connect the remagnetizing event to the regional dolomitization of these rocks, the evidence is still equivocal. It is quite likely that there is a connection between the two events but concrete proof of the physical and chemical relationship between the ChRM carriers
and the carbonate fraction remains elusive. McNeill and Kirschvink (1993) have shown for carbonates in the Bahama platform that early dolomitization helps "lock-in" clusters of pseudosingle domain magnetite, thus preserving the ChRM through the latter stages of diagenesis. Sites which were not dolomitized early on or which were dedolomitized at a later time in their diagenetic history did not preserve the earlier magnetizations. If this behaviour can be extended to the carbonates of the midcontinent, then any post-Permian dolomitization in the area of those sites that retained the A component would be restricted to a minor local event because the physical changes in carbonates associated with the process of dolomitization would destroy any preexisting ChRM direction.

Lu et al. (1990) and Jackson et al. (1992) have noted the general coincidence between basement structural highs and the preservation of ancient magnetizations. The same observation can be made for the geographic distribution of the A component. It is best preserved around the arch of the Nashville dome and Eastern edge of the Pascola arch. It is notably absent in the Illinois basin.

2.5.2 C Component

The C component was the only ChRM component observed in all sites from five quarries (table 2.2)(figure 2.9, solid circles). Three other quarries (figure 2.9, shaded circles) had one or more sites which were
<table>
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**Table 2.2**

Site mean statistics from those sites retaining the C component. Quarry abbreviations as in figure 2.1; Other abbreviations as in table 2.1.
unstable or yielded hybrid A-C components with the remaining sites from those three quarries retaining C as their ChRM. The combined population of the 19 site mean directions from the eight quarries that retained C as their ChRM yields a unit mean of D=355.6°, I=64.3°, α₉⁵=3.7°, κ=71, using Fisher (1953) statistics (figure 2.10). This corresponds to a pole position at 102.7°W and 80.8°N (δp=5.1°, δm=6.4°)(figure 2.11a) which is statistically distinct at the 95% confidence level from the Jurassic pole found by Symons (1994) in the fluorspar deposits of southern Illinois but statistically the same as that of the PEMF. However, because these ChRM directions were isolated in the 20 to 150 mT range, it is difficult to attribute their origin to viscous overprinting. Thermal overprinting can be ruled out because the ChRM direction survives laboratory heating of the specimens to over \( 500 \) °C. This equates to more than 100 Ma at 300°C (Pullaiah et al., 1975) which is inconsistent with the rock types, conodont alteration indices and fluid inclusion data for the area (Sangster et al., 1994). The only remaining alternative is that the C component sites were subjected to recurring fluid interaction that caused the complete destruction of the Permian magnetite, followed by reprecipitation of PSD to SD magnetite to yield a late Tertiary or modern C ChRM.

There are two implications to the fact that the C population ChRM direction is exclusively normal polarity and aligns extremely close to the PEMF direction rather than the long-term averaged axial dipole direction. It
suggests, first, that the C component was acquired over a very short period of
time and, second, that the magnetization is very young, having formed
within the present or latest normal polarity chron (1N) which restricts its age
to 0.78 Ma or younger (Cande and Kent, 1995). Estimates of the length of
time required to average out the effects of secular variation range from as
low as 2000 years (Tarling, 1983) to no greater than 27,000 years (Opdyke,
1972) which indicates that the C component may be significantly younger
than 0.78 Ma.

There is very little independent evidence for such a recent event in
the immediate district. The presumed Tertiary uplift of the Ozark dome to
the southwest could have remobilized formational fluids but this has not
been demonstrated geochemically. Also, if the fluids coming off the Ozark
dome were near the surface, one would expect them to produce hematite
rather than magnetite. Alternatively, surficial runoff from the last glacial
advance may have provided the fluid driving mechanism resulting in the C
component. The Michigan lobe of the Laurentide ice sheet did advance as far
south as mid Illinois at about 18 Ka (Dyke and Prest, 1987) but its southern
margin was still at least 200 km north of the area where the C component
was found.

Paleomagnetic studies by Lu et al. (1990) and Jackson et al. (1993)
have indicated the presence of "post-Kiaman" or "unstable" components in
the same general localities of the C component from this study. Since C was
isolated at magnetization intensities very near the noise level of our magnetometer, and our magnetometer is one of the most sensitive, it is possible that their “unstable” and “post-Kiaman” components are equivalent to the C ChRM.

Brannon et al. (1995) have obtained an 8 Ma U/Pb isochron for late stage calcite from the Permian ores and remagnetized host rocks of the northern Arkansas - Tri-State district (Pan et al., 1990) to the west of the Ozark dome. Brannon et al. (1995) also found that late stage calcite from the CT district gives a 39 Ma isochron. Their results show that mineral-forming events, and thus fluid migration events, continued long after the main ore-stage mineralizing event, perhaps with the fluids following the same preferred pathways.

2.6 Conclusions

Paleomagnetic analysis of carbonates from the Tennessee, Kentucky and Illinois area of the midcontinent yields two distinct magnetization directions. The A or oldest component is a late Paleozoic remagnetization that corresponds to the end of the “Kiaman” reversed superchron. The distribution of this population, as defined by Bingham (1974) statistics, suggests that it was acquired over a period of time of at least 8 Ma and possibly much longer.
The C ChRM, a much younger magnetization, is probably related to a further remagnetization from remobilization of magnetite by fluids within the last several thousand years. The C ChRM was isolated in rocks with an extremely weak magnetic moment and it is generally associated with basement lows.

No magnetization was found in this study that resembled the Jurassic magnetization (figure 2.10a) observed by Symons (1994) even though some of the quarries were within a few kilometers of fluorspar mines of the IK district (figure 2.9). This leads to the conclusion that the Rosiclaire fluorite mineralization event was restricted to the immediate vicinity of the veins and fractures containing the fluorite or that the evidence of this Jurassic event was obliterated by the late Tertiary event that caused the C ChRM except where the host rock was most intensely affected by the mineralizing fluids.
Chapter 3

Paleomagnetism and genesis of the
Elmwood-Gordonsville Mississippi Valley-type ore
deposit, central Tennessee

3.1 Introduction

The Elmwood-Gordonsville Mississippi Valley-type (MVT) deposit in
the central Tennessee (CT) district has been mined for over 25 years. During
that time numerous studies have attempted to understand the genesis of the
ore. However, one of the most critical pieces of information, the precise
timing of the mineralizing event, has remained elusive due to the absence of
suitable isotopic minerals and the general absence of post-ore crosscutting
relationships – a problem common to many MVT districts as previously noted
(Sangster, 1986). This study provides a date and an estimate for the length of
the mineralizing process in the CT district.
3.2 Geology

The CT MVT district is located on the northeast edge of the Nashville dome, a structural high of proto-Paleozoic origin, at the southern end of the Cincinnati arch (figure 3.1). The mineralization is hosted by carbonate rocks of the Ordovician Knox supergroup. The two major deposits, the Elmwood which opened in 1975 and the Gordonsville which opened in 1982, have been described in detail by Gaylord and Briskey (1983). They are only a few kilometres apart and have now been joined underground to form the Elmwood-Gordonsville mine under the operation of the Savage Mining Company.

Both Kyle (1976) and Gaylord and Briskey (1983) have provided detailed paragenetic sequences for the mineralization in the Elmwood-Gordonsville mine. The mineralization consists of open-space filling of pre-existing collapse breccias by sphalerite with minor barite, galena, fluorite and calcite. Dark reddish-brown sphalerite dominates with lesser amounts of intergrown light brown sphalerite. A local coarse-grained hydrothermal or sparry dolomite halo surrounds the ore and is associated with the process of mineralization. It can be easily distinguished from the fine-grained regional dolostone of the Knox supergroup. Regional dolomitization occurred in several stages, all of which appear to predate the mineralization (Montañez, 1995).
Figure 3.1

Regional map of the midcontinent of U.S.A. District abbreviations: CT=central Tennessee zinc; IK=Illinois Kentucky fluorspar; ET=east Tennessee zinc; CK=central Kentucky; SeM=southeast Missouri; NA=northern Arkansas; TS=Tri-State; CM=central Missouri.
It is now generally accepted that the driving force for MVT ore genesis in the CT district is nearby orogenic activity (Garven and Freeze, 1984a, b). In the model, Permian uplift associated with the Alleghanian-Ouachitan orogeny formed the Appalachian and Ouachita mountain chains, and the coeval fluid flow is presumed to have been responsible for the MVT mineralization here and elsewhere in the midcontinent of the United States (Oliver, 1986). With uplift the metal-bearing brines were driven out of the Black Warrior basin by changes in hydraulic head associated with orogenesis (Oliver, 1986, 1992), as previously described, and they migrated into the foreland bulge to deposit MVT mineralization.

Paleozoic strata in the CT region have slightly-elevated fluid inclusion temperatures (50 to 90°C) and conodont alteration indices (CIA=1.5) (Sangster et al., 1994) with respect to their expected burial temperatures due to the effect of warm brines passing through them. Fluid inclusions from the ore minerals show that the temperature of the mineralizing fluid was about 100° to 140°C (Gratz and Misra, 1987; Sangster et al., 1994). Thus the ore fluid was about 50°C hotter than any regionally extensive brine in its host strata. This contrast also favors the theory that at least two brines or two phases of brine migration must have occurred in the CT ore district.
There have been several attempts to place time constraints on the formation of the CT ores. The middle Ordovician age of the host-rocks provides a maximum age on ore genesis. Vein-type mineralization that is similar to the ores, is found in rocks as young as Early Mississippian (Jewell, 1947; Gratz and Misra, 1987). If the mineralization in the veins is coeval to the ores then the ores must postdate Early Mississippian. Kyle (1976) noticed that early breccias are barren while later breccias are often cemented with a matrix of MVT mineralization. This suggests that the mineralization postdates all of the post-Knox unconformity brecciation in the area. Estimates for the time of the brecciation range from slightly post Middle Ordovician (Hill and Wedow, 1971) to Late Devonian when the overlying Chattanooga shale was deposited (Hershey and Maher, 1985). Brannon et al. (1995) have obtained a 39 Ma U/Pb isochron from post-ore calcite so that the ore-stage mineralization must be older than 39 Ma. They also obtained a Th/Pb age of 260 +/- 42 Ma on main stage calcite which argues that the ore is roughly late Paleozoic. Fission track analysis of ore-stage fluorite, on the other hand, gives a 358 +/- 13 Ma date (V. M. Harder, unpublished data, pers. comm.) which suggests that the mineralization is mid Paleozoic in age.

3.3 Relationships to Adjacent Districts

The CT district has been associated with the ET MVT district, which is about 200 km to the east (figure 3.1), based on both mineralogical and
structural evidence (Kyle, 1976; Gaylord and Briskey, 1983). In both districts the main ore mineral is sphalerite and the host rocks are shallow-water dolostones of the lower to middle Mascot and the upper Kingsport members of the Knox supergroup. The process of ore genesis in both districts was the simple infilling of preexisting collapse breccias. However, the ET district is located within the fold-and-thrust belt of the Appalachians and it is structurally far more complex than the CT district which is in the foreland bulge where the sediments are nearly flat lying (dip <3°). The age of the ET deposits has been the subject of much debate and remains equivocal. A wide variety of structural and isotopic studies have ascribed mineralization ages ranging from pre-Middle Ordovician based on the restricted stratigraphy of the zinc occurrences (Hoagland et al., 1965; Crawford and Hoagland, 1968), to late Paleozoic (322 to 278 Ma) based on $^{40}$Ar/$^{39}$Ar isotopes in both authigenic feldspar (Hearn et al. 1987) and illite-smectite (Elliot and Aronson, 1987). Discordant Rb/Sr ages on sphalerite of 377 +/- 29 and 347 +/- 20 Ma have been obtained by Nakai et al (1990, 1993). A reversed polarity, late Paleozoic, Kiaman reversed superchron magnetization has been obtained from ores and host rocks in the ET district. It has been interpreted as a post-ore chemical remagnetization related to regional brine flow during the late Paleozoic (Bachtadse et al., 1987), similar to many other Paleozoic rocks from the interior craton (McCabe and Elmore, 1989). However it is possible that the magnetization in ET is coeval to mineralization because several
other paleomagnetic studies on other MVT ore deposits in the midcontinent have revealed late Paleozoic magnetizations that are coeval with ore genesis (Pan et al., 1990; Symons and Sangster, 1991; Elmore et al., 1993).

The CT district has also been compared genetically to the IK fluor spar deposits that are about 150 km to the northwest and to the Central Kentucky MVT deposits that are about 300 km to the northeast (figure 3.1). Structural similarities include the relationships to domes, namely the Hicks dome in Illinois and the Nashville dome in Tennessee, and to broad regional arches, namely the Tolu arch in Illinois and the Cincinnati arch in both Kentucky and Tennessee. Hayes and Anderson (1992) have observed an apparent correlation of fluorite zoning between all three districts. If correct, this indicates the presence of an extremely large regional brine flow through >5 x 10⁴ km³ of rock. Fluorite mineralization in the IK district is also seen in lower Permian igneous rocks which places an upper age limit on the mineralization. If the presence of the regionwide brine is valid, then the same age limit can be placed on the CT district. Hayes and Anderson (1992) propose a southern Appalachian provenance for this brine which could possibly tie in the ET deposits geographically and thus temporally. Fluorite from the IK district yields a Nd/Sm ratio 277 +/- 16 Ma (Chesley et al, 1994) that is coeval to hornblende and biotite ⁴⁰Ar/³⁹Ar dates of 271.7 +/- 1.4 and 272.7 +/- 1.4 Ma respectively from the alkalic intrusion at the core of Hicks dome (Snee and Hayes, 1992). This suggests that the fluor spar ores are a
local product of the formation of Hicks dome and, therefore, they cannot be related to either the CT or ET ore districts. A recent paleomagnetic study (Symons, 1994) revealed a Jurassic age for the magnetization in the fluorspar ores of the IK district which is in direct disagreement with any of the proposed ages for the ET magnetization (Bachtadse et al., 1987). This also argues against any temporal connection between the mining districts of ET and IK, which in turn questions each of their relationships to the CT district as well as the validity of the existence of a regionwide brine as proposed by Hayes and Anderson (1992).

3.4 Collection and Measurement Methods

Sites 1 to 26 were obtained during the summer of 1991 with 21 from within the Elmwood-Gordonsville mine and five at the Vulcan Materials quarry just south of Lebanon, Tennessee, less than 20 km from the mine. Sites 31 to 43 were collected from within the mine during the summer of 1993. Four to six oriented blocks of hand sample size were taken at each site in the mine. An attempt was made to cover all the lithological and mineralization types. Massive and disseminated ores as well as ore-stage accessory minerals such as sparry or hydrothermal dolomite were collected in addition to altered and unaltered host rock. Four sites of sphalerite, barite and fluorite crystals were also obtained. Three sites consisting of eight or nine oriented breccia clasts each were obtained for breccia tests. One site of
fine grained dolostone was collected from the top of the mine ramp. At the quarry, four oriented blocks per site were obtained from five sites in limestones spanning about 10 metres of strata.

All blocks were drilled in the laboratory, yielding 10 to 15 core specimens per site for a total collection of about 500 specimens. The collection was stored for up to one year in the shielded room thus significantly reducing the effects of viscous remanence components.

All natural remanence magnetization (RM) measurements were made on the CTF cryogenic magnetometer. Four pilot specimens, two AF and two thermal, were selected from each site for detailed demagnetization. AF pilots were treated in 12 to 14 steps to a peak field of 150 mT. Thermal pilots were treated in 10 to 12 steps to 600°C where possible. Based on the results of the pilots the remaining specimens were treated in four or five AF fields ranging from 20 to 140 mT. Several additional specimens, representative of the collection as a whole, were selected for SIRM analysis to better define their magnetic mineral carriers. SIRM experiments were carried out in the paleomagnetic laboratory at MacMaster University in Hamilton, Ontario, using a Molspin magnetometer as well as on the CTF magnetometer at the University of Windsor.
3.5 Results

In most cases AF demagnetization yielded better results than thermal demagnetization. Treatments to as high as 150 mT resulted in consistent directional data with significant decay of the NRM, particularly from the unaltered host rocks. There was a direct relationship between remanence stability, NRM intensity and lithology. Barren host-rock limestones and dolostones (figure 3.2) were generally more reliable and more intensely magnetized than either disseminated (figure 3.3) or massive ores. There are two possible causes for this observation. It could reflect differences in the inherent stability characteristics of the magnetic carriers or it could be a product of equipment limitations. NRM intensities ranged from the mid $10^{-4}$ to low $10^{-6}$ A/m range with limestones in the higher end and ore-bearing sites in the lower end of the intensity range. Many of the more weakly magnetized specimens were close to or at the $\sim 4.0 \times 10^{-6}$ A/m sensitivity limit of the magnetometer. This resulted in a significant increase in the scatter of directions upon demagnetization of the weak specimens and the rejection of an unusually high percentage of their data ($\sim 50\%$) due to lack of signal. AF pilots that were stable showed significant decay of the NRM on demagnetization. This provides evidence that hematite is not present as a remanence carrier because it is virtually unaffected by AF fields of 150 mT or less.
Figure 3.2

Normalized orthogonal demagnetization plots of AF pilot specimens from limestones in the quarry. Circles are projections in the N-E-S-W plane and triangles are projections in the N-Down-S-Up plane. Treatments are in milliTesla and NRM intensities are in A/m.
Figure 3.3

Normalized orthogonal demagnetization plots of AF pilot specimens from four sites that were either mineralized or visibly altered by hydrothermal dolomite. Symbols as in figure 3.2.
Thermal pilots were less efficient at identifying remanence components. The problem of sulphides in the host rocks generating magnetic minerals in the oven at temperatures in excess of 400°C was discussed in the previous chapter. Intensity jumps in excess of 10x, coupled with inconsistent directional data were often observed in the 400 to 500°C range (figure 3.5b). As a result most thermal pilots from ore sites had to be abandoned by 500°C. Significant intensity drops in the unblocking temperature ranges of goethite (80-120°C) or pyrrhotite (280-320°C) were not observed which excludes them as possible carriers of the remanence. The relative contributions of magnetite and hematite, based on thermal demagnetization of the NRM alone, is impossible to determine because the specimens were abandoned before reaching their respective 585°C and 670°C unblocking temperatures. However the behaviour of the AF pilots indicates that hematite is not an important contributor to the remanence. The directions observed in the thermal pilots up to as high as 500°C were concordant with the directions of the corresponding AF pilots.

Example orthogonal demagnetization plots for specimens from the quarry (figure 3.2) show that the same ChRM direction was observed in all five sites at the quarry. This direction was south-southeast and equatorial after the removal by 30 to 50 mT (figure 3.2a, b) of a steeply-downward viscous component that presumably records the PEMF. The antiparallel,
Normalized orthogonal demagnetization plots of AF pilot specimens from the mine. Notice that the two breccia sites display behaviours similar to the disseminated ore and dolomites. Symbols as in figure 3.2.
north-northwest and equatorial direction was also observed, but less often (figure 3.2c, d).

Figure 3.3 shows example orthogonal demagnetization plots for four specimens of ore from within the mine. Again a south-southeast and equatorial ChRM was observed after removal of a steeply downward viscous component (figure 3.3a, b, c). As in the quarry, specimens with the opposing polarity were observed (figure 3.3d) but with less frequency.

Orthogonal demagnetization plots for an additional ore specimen (figure 3.4a), a clast from an early breccia (figure 3.4b), a dolomite remote from any mineralization (figure 3.4c), and a clast from a late breccia (figure 3.4d) show that the south-southeast ChRM was observed in all of these sites. More importantly it shows that the magnetization retained in the clasts of these two breccias postdates brecciation in the central Tennessee district but is coeval with the magnetization from the ores (table 3.1). The third breccia site did not retain stable directions at the specimen level and, therefore, neither a positive nor negative result can be reported for it.

Figure 3.5 shows example orthogonal plots for specimens that were demagnetized thermally. Figure 3.5a shows one of the few specimens that was not adversely affected by heating in the oven. It is completely demagnetized by the 585°C Curie temperature of magnetite. Figures 3.5b and 3.5c show the more common behaviour of specimens subjected to thermal demagnetization. Notice that the direction becomes erratic by about 400 to
<table>
<thead>
<tr>
<th>Site</th>
<th>Rock Type</th>
<th>( N_M - N_R - N_N )</th>
<th>Dec.</th>
<th>Inc.</th>
<th>( \alpha_{95} )</th>
<th>( \kappa )</th>
</tr>
</thead>
<tbody>
<tr>
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<td>disseminated sphalerite</td>
<td>12-12-0</td>
<td>174.4</td>
<td>2.7</td>
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<td>164.8</td>
<td>2.8</td>
<td>12.2</td>
<td>14</td>
</tr>
<tr>
<td>7</td>
<td>disseminated barite</td>
<td>12-3-3</td>
<td>166.1</td>
<td>3.2</td>
<td>18.9</td>
<td>14</td>
</tr>
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<td>chert nodules</td>
<td>12-4-2</td>
<td>169.4</td>
<td>0.5</td>
<td>16.8</td>
<td>17</td>
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<td>21</td>
<td>early breccia</td>
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<td>4.7</td>
<td>40</td>
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<td>14</td>
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<td>21</td>
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<tr>
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<td>166.3</td>
<td>5.0</td>
<td>4.8</td>
<td>161</td>
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<tr>
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<td>9-3-0</td>
<td>167.7</td>
<td>1.4</td>
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<td>27</td>
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<td>12.0</td>
<td>26.3</td>
<td>13</td>
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<tr>
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<td>hydrothermal dolomite</td>
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<td>168.4</td>
<td>1.1</td>
<td>9.6</td>
<td>30</td>
</tr>
<tr>
<td>all 14</td>
<td>ore sites</td>
<td>197-97-14</td>
<td>168.8</td>
<td>3.7</td>
<td>3.5</td>
<td>128</td>
</tr>
</tbody>
</table>

Table 3.1

Site mean statistics from those sites retaining ore stage magnetization. \( N_M - N_R - N_N \) are the number of specimens measured, those with reversed components and those with normal components respectively; Dec and Inc are the declination and inclination; \( \alpha_{95} \) is the cone of 95% confidence and \( \kappa \) is the precision parameter (Fisher, 1953).
Figure 3.5

Normalized orthogonal demagnetization plots of thermal pilot specimens from the mine (a, b, c) and quarry (d). Treatments are in °C, other symbols as in figure 3.2.
450°C although the directional behaviour of these pilots below that temperature is similar to that of the AF pilots. Figure 3.5d shows a specimen that appears to exhibit both polarities. Between 300 and 450°C, it swings from a shallowly-upward direction to the north to an antiparallel, shallowly-downward direction to the south. This behaviour was not observed in any other specimens so the recording of a reversal cannot be verified.

The results of the SIRM analyses are shown in figure 3.6. All curves fall between the type curves for single domain (SD) to pseudosingle domain (PSD) magnetite both on acquisition and decay except for: a) two specimens with acquisition curves below the SD type curve which indicates that a minor hematite component may have contributed to the signal in these two specimens; and b) two specimens with decay curves below the PSD type curve which indicates that multidomain magnetite is also present. No evidence of hematite was seen in demagnetization of the NRM in any of the AF or thermal pilots which suggests that the hematite in these two SIRM specimens is probably the product of recent surficial weathering. There was no bias in SIRM behaviour between specimens of differing lithology or mineralization type except that massive sphalerite, barite and fluorite samples were too weakly magnetized to yield a useful signal, even after direct field magnetization at 900 mT, so only disseminated ore specimens are included. These results, combined with the specimen behaviour on both AF and thermal demagnetization of the NRM, show that magnetite is the sole
Figure 3.6

Saturation isothermal remanence saturation curves of example specimens representative of the collection as a whole. Standard acquisition (a) and decay curves (b) show normalized intensities as a function of treatment field in milliTesla (mT). Dashed lines are the theoretical ranges for multi- (MD), pseudosingle (PSD), single (SD) domain magnetite as well as fine (FH) and coarse (CH) grained hematite.
carrier of the ChRM and that there is no difference in the magnetic characteristics of the ore-bearing specimens when compared to their host rocks.

Figure 3.7 shows the Cisowski (1981) crossover points for the same set of SIRM specimens. These are obtained by plotting the normalized SIRM acquisition and decay data for a single specimen on the same horizontal axis and noting the intersection point of the two curves. The theoretical range for crossovers for PSD to SD magnetite is from 30 to 60 mT and below 0.5 of the initial intensity. In general, as 0.5 is approached, directional stability increases as the grain size and/or degree of interaction between adjacent magnetic domains decreases. All specimens with one exception plot within this field which supports the conclusion that SD to PSD magnetite is the carrier of the ChRM. This conclusion agrees with the rock magnetic studies on remagnetized Paleozoic carbonates in chapter two and from elsewhere in the midcontinent (e.g. Lu et al., 1990; Jackson et al., 1992).

Specimen data were again analyzed using the principal component method of Kirschvink (1980) with directions having mean deviation angles of greater than 15° being rejected. Directions were only anchored to the origin when it was visually evident that the unanchored direction was trending towards the origin over a significant portion (>20%) of the demagnetization spectrum. In all cases the difference between the anchored and unanchored vectors was less than 10° of arc. End point directions that passed these
Figure 3.7

Cisowski (1981) crossover points of SIRM test specimens. Dashed line is the theoretical upper limit for pure single domain magnetite.
criteria were obtained from 20 sites which, in general, were the most intensely magnetized sites in the collection. After removal of a steeply-downward viscous component by about 30 mT, these 20 sites isolated a high coercivity ChRM that is directed towards the south-southeast and almost horizontal. Site means were generated first using the statistical methods of Fisher (1953) (figure 3.8) (table 3.1, table 3.2).

The remaining sites in the collection were either too weakly magnetized to be reliably measured or did not pass the screening criteria for stability. In particular, all of the massively mineralized and single crystal specimens were too weakly magnetized to be measured and thus were abandoned. This meant that several sites containing 100% MVT mineralization were immediately lost to further analysis.

3.6 Discussion

3.6.1 Host Rocks

Four of the five sites in the quarry form a tight population of ChRM directions in the south-southeast quadrant of a stereonet (figure 3.8) after reversing the directions of specimens that retained the opposing polarity (table 3.2). One site of fine-grained regional dolostone from the top of the access ramp to the mine, site 18, is remote from the ore. It is mapped as a regional dolostone, distinct from the ore-stage sparry dolomite, by the mine
Figure 3.8

Stereonet projection of the site means (circles and squares) and overall mean (star) for the ore and host rock populations. The two breccia tests (sites 22 and 23) are noted by the squares. The host rock mean includes site 13.
<table>
<thead>
<tr>
<th>Site</th>
<th>Rock Type</th>
<th>$N_M - N_R - N_N$</th>
<th>Dec.</th>
<th>Inc.</th>
<th>$\alpha_{95}$</th>
<th>$\kappa$</th>
</tr>
</thead>
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<tr>
<td>13</td>
<td>limestone</td>
<td>13-6-0</td>
<td>135.0</td>
<td>-1.6</td>
<td>25.1</td>
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<td>14</td>
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<td>95</td>
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<tr>
<td>17</td>
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<td>152.8</td>
<td>-1.0</td>
<td>6.6</td>
<td>37</td>
</tr>
<tr>
<td>18</td>
<td>fine grained dolostone</td>
<td>11-7-1</td>
<td>152.3</td>
<td>5.3</td>
<td>15.6</td>
<td>16</td>
</tr>
<tr>
<td>all 6</td>
<td>hosts</td>
<td>75-58-6</td>
<td>149.2</td>
<td>1.2</td>
<td>5.1</td>
<td>143</td>
</tr>
<tr>
<td>excl. #13</td>
<td>hosts</td>
<td>62-52-6</td>
<td>152.1</td>
<td>1.3</td>
<td>2.6</td>
<td>881</td>
</tr>
</tbody>
</table>

**Table 3.2**

Site mean statistics from those sites retaining host rock magnetization. Abbreviations as in table 3.1.
staff, and its mean ChRM direction fits within the limestone population from
the quarry. Thus its data can be validly combined with the data from the
quarry. The remanence direction for the most aberrant quarry site, site 13, is
about 15° away from these five sites and as such it is questionable as to
whether or not it should be included in this population. Taken together these
six sites give a mean of declination (D)= 149.2°, inclination (I)= 1.2°, radius of
95% confidence (α95) = 5.1°, precision parameter (κ)=143 (figure 3.8b) which
corresponds to a pole position, after reversing to its antipodal position, at
138.9°E, 43.6°N (δp=3.1°, δm=6.2°)(Fisher, 1953). If the direction for the
deviant site 13 is rejected, the five remaining sites have a population mean of
D= 152.1°, I= 1.3°, radius α95 = 2.6°, κ=881 which shifts the pole position
about 4° to 135.5°E, 44.9°N (δp=1.3°, δm=2.6°) (Fisher, 1953) and provides a
better fit to the APW path (figure 3.9, pole “hosts”). This direction is similar
to the Kiaman reversed superchron (~320-250 Ma) magnetizations of
McCabe et al (1989) who report a pole position of 127.9° E, 47.9° N, A95=5.2°
(figure 3.9, pole “Tn2”) as well as the A component of this thesis which gives
a pole position of 126.9°E, 46.1°N (δp=1.5°, δm=3.0°)(figure 3.9, pole "Tn3").
Both the study of McCabe et al (1989) and of this thesis were on surficial
exposures of Paleozoic carbonates exposed in quarries or road cuts along
transects across Tennessee. The results of these two studies plus the five
sites presented here provide an accurate regional paleomagnetic signature
for the Paleozoic strata of Tennessee, showing that virtually all of the
North American Paleozoic apparent polar wander path showing the results of this section (hosts and ores) compared to published reference poles for Paleozoic carbonates of Tennessee by Bachtadse et al. (1987, Tn1), McCabe et al. (1989, Tn2) as well as the A component (Tn3). All confidence estimates are 95% ($\Delta_{95}$) and are by Fisher (1953). The grey “hosts” mean includes all six sites while the black mean excludes the aberrant site 13.
carbonates were uniformly remagnetized in early Permian time over an area of at least 5 x 10^4 km^2. These data suggest that the entire package of Paleozoic carbonates may have acted as a giant aquifer during the Permian.

3.6.2 Ores

Only 14 of the 33 sites from ore stage material inside the Elmwood-Gordonsville mine retained consistent ChRM directions. Of the 19 sites that did not retain ChRM's, 14 were simply too weakly magnetized to be measured. Five other sites, including one of the breccia sites, were unstable and failed to retain coherent ChRM's. Of the 14 remaining sites, two consisted of clasts from early (site 21) and late (site 22) breccias. Since both of these sites retained ChRM's that agreed with those from the ores, it is clear that the magnetization postdates brecciation in the mine. These two sites were combined with the remaining 12 ore sites to form a population mean from the mine of D=168.8°, I=3.7°, α_95 =3.5°, κ=128 (Fisher, 1953). It corresponds to a pole position, again after reversal, at 111.9°E, 50.5°N (δp=1.7°, δm=3.5°)(Fisher, 1953) (figure 3.9, pole "ores") which falls on the Permo-Triassic boundary of the North American APW path. It indicates an age for the magnetization of 245 +/- 15 Ma. This age determination agrees favourably with the age restrictions placed on the ore deposit from brecciation (Hill and Wedow, 1971; Hershey and Maher, 1985), stratigraphic occurrences (Jewell, 1947; Gratz and Misra, 1987) and U/Pb age dating on
post-ore calcite (Brannon et al., 1995) (figure 3.10). It is also in direct
agreement with the main ore stage Th/Pb date of 260 +/- 42 Ma (Brannon et
al., 1995). The paleomagnetic age also fits well with the late Paleozoic
basinal fluid expulsion model (Oliver, 1986, 1992) that has been applied to
other MVT ore districts in the midcontinent. The only conflicting result to the
paleomagnetic age is the unpublished fluorite fission track age of 358 +/- 13
Ma (V. M. Harder, pers. comm.).

The CT data can be compared to the paleomagnetic data obtained
from the ET mining district by Bachtadse et al. (1987). They argued that
their data recorded a late Paleozoic, post-ore, remagnetization event. Their
results agree at the 95% confidence level with the host rock results discussed
earlier (figure 3.9, pole "Tn1") but they differ from the ore magnetization
direction found in CT district. This divergence provides evidence that the ore
in ET is not coeval to that of CT and, therefore, probably not the product of
the same fluid event as the ore in CT. As discussed in section 3.3, the late
Paleozoic $^{40}$Ar/$^{39}$Ar results for ET are probably recording the onset of the
fluid expulsion event that is responsible for the CT ore genesis while the
older Rb/Sr ages are probably recording the true date of MVT ore genesis in
ET.

The CT ore pole also differs from the pole found for the fluor spar
ores of the IK mining district (Symons, 1994). If the Nd/Sm dating for the IK
district is correct, then it is a local event caused by hydrothermal circulation
Figure 3.10

Proposed ages for the central Tennessee MVT ore district according to the methods listed. References to these data can be found in the text. The arrows indicate that the age estimate is unrestricted in that direction. The dashed line is the proposed age range for brecciation which is inferred to have been completed prior to mineralization.
stemming from Hicks dome. If the paleomagnetism is correct, then the IK flourspar district is Jurassic in age and probably related to rifting. In either scenario there is no reason to believe that IK district is genetically related to the CT district.

3.6.3 Comparison of ore and host poles

The direction obtained from the ores is similar to, but statistically distinct at the 95% confidence level from, the regional direction obtained from host rocks throughout Tennessee (figure 3.9). Also the trend of the bias in the dispersion of these two populations differs when described by Bingham (1974) statistics. Unlike Fisher (1953), the Bingham ellipse is free to elongate parallel to the trend of elongation in the data at the site mean level, as described in chapter one and more fully in Lewchuk and Symons (1995)(appendix C). When analyzed using the method of Bingham (1974) the resulting confidence ellipses are $\alpha_{1,2} = 3.5^\circ$, $\alpha_{1,3} = 1.9^\circ$, azimuth $= 124^\circ$ for the A component and $\alpha_{1,2} = 4.0^\circ$, $\alpha_{1,3} = 2.1^\circ$, azimuth $= 69^\circ$ for the B component. A closeup view of the site mean data for the B ChRM found in the hydrothermal or sparry dolomite and disseminated ore (figure 3.11a) and the site mean data for the A ChRM found in the regional host-rock dolostones (figure 3.11b) demonstrates the difference. Notice that the two ChRM populations differ in both direction and azimuth of elongation. Figure 3.11c shows both poles relative to the APW path of Van der Voo (1990). While both
Magnification of equatorially centered stereonet showing: a) the site means for the ore magnetization (circles); b) the A ChRM component from regional carbonate host rocks of the quarry transect and; c) the poles and their error ellipses superimposed on the North American apparent polar wander path. The lines in a and b represent the azimuth of elongation according to Bingham (1974).
poles are concordant to the Phanerozoic path. They are statistically distinct
from each other at the 95% confidence level with the host-rocks' pole in the
Early Permian and the ore pole straddling the Permo-Triassic boundary.
This provides a positive "chemical contact" test which indicates that the B
ChRM is primary and, therefore, coeval to the sulphide mineralization, even
though no magnetic signature was found in either massive or pure
crystalline mineralization. The NRM intensities of these rocks are so low
that only minute amounts of magnetite are required. Very small quantities of
magnetite have been observed in the paragenetic sequence of the Southeast
Missouri MVT district (Hagni, 1989), showing that trace amounts of
magnetite are capable of being formed in a sulphide-dominated environment.
Thus this magnetization is deemed to be recording the genesis of the ore at
about 245 Ma. More importantly, the elongation in both populations is in
general agreement with the trend of the APW path for the time period
indicated by their mean directions. This provides evidence that the
elongation is the product of APW during the interval over which the
magnetizations were acquired. Assuming this is true, the method of Lewchuk
and Symons (1995) can be used to estimate that ore of the CT district
acquired its magnetization over an interval of about 10 Ma. The length of
time estimated here is in general agreement with the theoretical calculation
of Garven and Freeze (1984a, b) based on numerical modeling of heat flow,
heat transport and mass transport with water-rock reactions.
3.7 Conclusions

The mineralization in the CT district was formed over a period of about 10 Ma in latest Permian to earliest Triassic time. Ore magnetization is clearly several million years younger than regional host rock remagnetization which provides a positive “chemical contact” test that indicates that the ore magnetization is primary. The magnetization, and thus mineralization, in CT is unrelated to the IK district fluorspar ores. Thus the single regional brine model of Hayes and Anderson (1992) is not supported. The mineralization is also unrelated to the genesis of the MVT zinc ores in the ET district. They are most likely related to an earlier orogenic episode. The Permo-Triassic age for the CT MVT ore district fits well with the $^{40}\text{Ar}/^{39}\text{Ar}$ and paleomagnetic ages obtained from the ET district, but not with its Rb/Sr ages. Therefore the $^{40}\text{Ar}/^{39}\text{Ar}$ dates in ET are probably recording the Permian brine migration that regionally remagnetized the Paleozoic strata of Tennessee while the Rb/Sr ages may be recording mineralization.
Chapter 4

Summary

4.1 Regional Transect

Paleomagnetic analysis of 48 sites from Paleozoic carbonates collected at 16 quarries in Illinois, Kentucky and Tennessee isolated two characteristic remanence magnetization directions, named the A and C components, along a traverse extending from the CT MVT ore mining district to the IK fluorspar mining district. Both components are much younger than the Ordovician to Mississippian depositional ages of the rocks sampled. There is a trend from A to C on going from southeast to northwest along the transect.

The A component, found mostly in the southeastern or Tennessee part of the transect, has a mean of declination (D) = 157.2°, inclination (I) = 4.4°, radius of 95% confidence (α95) = 3.0°, precision parameter (κ) = 99. This corresponds to a pole position, after reversing to its antipodal position, at 126.9°E, 46.1°N (δp = 1.5°, δm = 3.0°) (Fisher, 1953) that falls on the Permian portion of the North American APW path. The A magnetization in these
rocks is statistically similar to "Kiaman" reversed interval overprint magnetizations that have been found by other researchers (see McCabe and Elmore, 1989) to be ubiquitous in Paleozoic carbonates of the midcontinent of the U.S.A. Although it has not been proven, this pervasive remagnetization is probably coeval with the regional pervasive dolomitization in the area as both are the product of post-diagenetic geochemical interaction with pore fluids. The A component was acquired over a period of no less than 6 Ma and possibly much longer, because the process of remagnetization itself may have obscured or overwritten the earlier part of the remagnetization record. When the A component is compared to the magnetization found for MVT ore in the CT district, it is clear that regional remagnetization of the Paleoozoic carbonates in the midcontinent preceded the precipitation of the ores.

The C component was seen in the northwestern part of the transect and is much younger than A. The mean direction for C of D=355.6°, I=64.3°, \( \alpha_{95}=3.7°, \kappa=71 \), using Fisher (1953) statistics corresponds to a pole position at 102.7°W and 80.8°N (\( \delta_p=5.1°, \delta_m=6.4° \)). This pole is virtually identical to the short-term PEMF pole showing that it was probably acquired over the last several hundred years or less. The magnetite retaining this component has a high coercivity so that the ChRM was too stable to be rejected as spontaneous viscous overprinting. Therefore an additional, as yet undefined, fluid remobilizing event must have occurred in this part of the transect. While the A component was generally found in carbonates overlying
basement highs, the C component appears to be associated with carbonates overlying basement lows.

No evidence was found of a magnetization matching the Jurassic direction seen previously in the fluor spar ores of the IK mining district. Therefore, the formation of the fluorite vein deposits was either a localized mineralization event that was restricted to the open spaces of the veins and adjacent host rocks alone or the paleomagnetic record of the fluorite mineralizing event was not preserved in the carbonate rocks away from ore in the district.

4.2 Central Tennessee Ore Deposits

Paleomagnetic analysis was conducted on more than 500 specimens from 43 sites in MVT ore, ore-stage hydrothermal dolomite and remote Paleozoic carbonate host rocks of the CT zinc mining district. While several sites obtained from pure crystal specimens of ore stage material proved to be too weakly magnetized to be successfully measured, disseminated ores, ore-stage hydrothermal dolomites and the remote host rocks all yielded useful paleomagnetic data that provided a paleomagnetic age for the genesis of the ores.

Rock magnetic experiments and the behaviour of the AF and thermal pilots on demagnetization of their NRM, indicate that fine grained,
non-interacting magnetite is the sole remanence carrier of consequence in the collection.

Five sites from a quarry only a few kilometres away as well as one site from the portal of the access ramp to the mine form a tight population of site mean ChRM directions that is statistically distinct at the 95% confidence level, using either Fisher (1953) or Bingham (1974) statistics, from the ore stage magnetization. These six sites give, statistically, the same direction as the A component sites from the carbonates of the regional transect as well as previously published regional host-rock directions for Paleozoic carbonates in Tennessee. This agreement indicates that all of these Paleozoic carbonate units were remagnetized in the Permian regardless of their stratigraphic position or carbonate geochemistry.

Paleomagnetic ChRM directional data from 14 sites, all located within the Elmwood-Gordonsville mine, that include partially-mineralized material, hydrothermal dolomite, early and late breccias were combined to form a B population mean direction for the ore-stage event of $D=168.8^\circ$, $I=3.7^\circ$, $\alpha_{95}=3.5^\circ$, $\kappa=128$. This corresponds to a pole position after reversal at $111.9^\circ E$, $50.5^\circ N$ ($\delta_p=1.7^\circ$, $\delta_m=3.5^\circ$) which falls on the Permo-Triassic boundary of the North American APW path, thus placing the age of the mineralization at about $245 \pm 15$ Ma.
The mineralization age given by the B or ore pole is slightly younger than the remagnetization age of the A or host rocks pole. This result proves by way of a "chemical contact" test that, although the host-rocks are remagnetized, the ore stage material must be younger and carrying a primary remanence, and thus it is recording the late Paleozoic genesis of mineralization in the CT district.

The results of this study indicate that the CT zinc district is not related to either the nearby ET zinc or IK fluorspar districts despite similarities in mineralization and stratigraphic control. Therefore, the apparent region-wide correlation of fluorite zonation seen by Hayes and Anderson (1992) is not supported.

The elongate distribution of the B component site means for the ore-stage material can be described using Bingham (1974) statistics. They indicate that the elongation is the product of APW during the process of "locking-in" of the magnetization as the deposit was formed over a period of about 10 million years. This estimate for the duration of mineralization is consistent with theoretical models based on geochemical constraints as proposed by Anderson and Macqueen (1982) as well as Garven and Freeze (1984a, b).
4.3 Significance for MVT ore genesis

Although new data are provided for only one MVT ore district, the application of a new statistical method to this data as well as to previously published data has several consequences for models of the genesis of MVT ore deposits in general.

The first conclusion is that there is a general relationship between the timing of host-rock remagnetization and ore formation. In all of the MVT districts studied paleomagnetically thus far, the host rocks were remagnetized shortly before the ore-forming event even though the host rocks were commonly in place for as long as three hundred million years prior to the precipitation of the ore deposits. This argues for the existence of two fluids, an early, metal-poor dolomitizing brine, possibly bearing sulphur, and a late nondolomitizing, metal-rich MVT mineralizing brine. The fact that a two-stage event was seen paleomagnetically in five different “carbonate platform-orogenic belt” pairs (see appendix C) suggests that this is a feature which is common to all MVT ore districts. This sequence has been suggested by Banner et al. (1988) for the southeast Missouri MVT district and Fontboté and Gorzawski (1990) for a MVT ore deposit in Peru.

It is certainly possible that this two-stage event is actually a continuum in which one brine that evolves through space and time from the early to the late composition. In this two-brine model it is possible, but not proven, that the earlier, base-metal barren, dolomitizing and/or
remagnetizing fluid was responsible for much of the karstification and that later ore fluids simply moved into the area and passively filled in the open spaces with little or no interaction with their host rocks. Thus, the MVT mineralizing brine does not have to be acidic enough to dissolve massive amounts of host rock material in order to generate the open spaces for the ore to form.

Another consequence of a two-brine model is that the sulphur and base metals are not required to be transported together. It is quite possible that the earlier brine transported the sulphur to the foreland bulge and then left it behind, most likely in the form of a trapped gas cap over petroleum. This would explain the general relationship between MVT ore deposits and oil producing areas (Oliver, 1992). It would also explain why the MVT ores form in foreland bulges and at stratigraphic pinchouts or reef structures, three features which commonly trap petroleum and natural gas (Anderson and Macqueen, 1982; Sangster, 1986).

Finally, the application of Bingham statistics to the available data show that, in general, MVT ore deposits take up to about 25 million years to develop. This is consistent with the previously published genetic and mathematical model of Garven and Freeze (1984a, b). They showed that the maximum likely life span of a hydraulic regime disturbed by a single overthrusting event would be about 200,000 years and that it would take several of these events to form an average MVT ore deposit. Anderson and
Macqueen (1982) used estimates of the relationship between flow-rate, fluid volumes and concentration change to show that, theoretically, 20 million tons of 5% ore could form in a million years. Since this represents a smaller much deposit than any of the ones studied here, the estimate of 25 million years given here is in excellent agreement.

4.4 Paleomagnetic Implications

There are two conclusions from this research that are important from the standpoint of paleomagnetism in general.

First, this research has been able to define the age and duration of multiple regional fluid flow events as well as to map their geographic extent. This shows that paleomagnetism could become a useful tool in problems associated with sedimentary basin analysis.

Second, Bingham statistics was successfully used to estimate the length of time over which the ore deposits formed. It is now clear that the method of Lewchuk and Symons (1995) is a viable technique for the estimation of the acquisition time of a magnetization assuming that the time frame in question is long enough to record APW. The data for this research were obtained from rocks that were in a tectonically stable regime so that local fault block rotations or the presence of unseen minor folds can reasonably be excluded. Given a similar set of geological circumstances there
is no reason why this statistical method cannot be applied to other geological phenomena.
Appendix A

The Coeur d’Alene Project

A paleomagnetic study was attempted on the argentiferous vein deposits of the Coeur d’Alene district of Idaho (figure A.1). The Coeur d’Alene district, including the Superior subdistrict of western Montana, is one of the premier silver mining districts in the world, having produced over one billion ounces of silver as well as eight million tons of lead, three million tons of zinc, 170,000 tons of copper and 500,000 ounces of gold (Leach et al, 1988). The ores are hosted by the Proterozoic Belt supergroup, a 20,000 m thick sequence of sedimentary rock that was deposited on the passive western margin of North America between 1750 and 850 Ma. Ore mineralization is found in veins that occur throughout the Belt basin but they are particularly well concentrated along the Lewis and Clark line (figure A.1). It is a major structural zone that has existed since Proterozoic time (Harrison, 1972). The only constraints on the age(s) of the veins are the Late Proterozoic age of the host rocks and the Late Cretaceous age of the Gem stock that cuts veins in
Regional map of the northwestern United States showing the Coeur d'Alene (CDA) and Superior (SD) silver mining districts along the Lewis and Clark line. The shaded area is the approximate extent of Belt supergroup rocks.
the northern part of the district. At the Sunshine mine in Coeur d'Alene, ore-grade silver veins cut 1100-1200 Ma uraninite veins. K/Ar determinations on sericite in post-ore hydrothermal veins have yielded ages of 829 +/- 43 (Bunker Hill mine), 876 +/- 43 (Lucky Friday mine), 447 +/- 25 (Galena mine) and 77 +/- 5 Ma (Sunshine Mine) (Leach et al, 1988). Zartman and Stacey (1971) provided model lead ages of 1200-1500 Ma for galena in the veins. They argued that it would be unlikely for these values to be preserved if there was any Phanerozoic remobilization of the lead. If one takes the model lead ages, the oldest of the post-ore K/Ar ages, and the age of the Lewis and Clark line, a convincing argument can be made for a Late Proterozoic age of vein mineralization. However, this is a very loose estimate with no direct dating evidence to support it. Recent unpublished K/Ar, ^40\text{Ar}/\ ^39\text{Ar} and lead isotope data suggests that there are two generations of veins, the silver-rich veins of Late Cretaceous age and the zinc-rich veins of Late Proterozoic age (A. H. Hofstra, pers. comm.). Given the ambiguity over the age of the veins, paleomagnetic methods were used in an attempt to date the veins. Two factors indicated that this might be successful: 1) galena-and zinc-rich ore deposits from the midcontinent, although of different genesis, had been successfully dated paleomagnetically; and, 2) the fluor spar vein-type deposits from Illinois had been successfully dated paleomagnetically. Negative factors included: 1) the complex tectonic and structural history of the western margin of North America; and, 2) the
possibility that the ores were Proterozoic in age, a period for which the
definition North American APW path is much less precise. As it turned out,
none of these played a role in the eventual failure of the experiment.

Three to six oriented hand samples were taken at each of 34 sites in
the Coeur d’Alene district. They included 14 sites from veins and two sites
from host rock in the Sunshine mine, four sites from veins in the Lucky
Friday/Gold Hunter mine, eight sites from the St. Regis formation of the Belt
supergroup, and six sites from the Cretaceous Gem stock. Oriented cores
were drilled from the hand samples in the laboratory and then sliced into
specimens to yield a collection of over 400 specimens. Two AF and two
thermal pilots were selected from each site for detailed demagnetization.

Initial results were very encouraging for the silver ore specimens.
Within-specimen directions deviated by 2° or less on subsequent
measurements and intensities varied by as little a 1% from NRM to 150 mT
of AF demagnetization in some cases. This normally implies that a very
stable, high coercivity component is present. However, there was little or no
consistency from one hand sample to another. The remanence intensity was
usually in the range of $10^{-3}$ A/m in the cryogenic magnetometer, yet the same
specimen failed to produce a signal when it was measured in a spinner
magnetometer with a low enough sensitivity threshold to measure such a
specimen. The problem was the presence of siderite ($FeCO_3$), an iron
carbonate gangue mineral that comprised as much as 90% of the vein
material. The siderite, with a Königsberger ratio (i.e. remanence magnetization/magnetic susceptibility) of $10^{-3}$, reacted to the trapped magnetic field in the superconducting coils to produce an induced magnetization that was as much as three orders of magnitude greater than the remanence magnetization.

Cryogenic magnetometers trap magnetic fields within their superconducting pickup coils as they are cooled to below the temperature required for superconductivity. This field remains constant as long as the coil remains in the superconducting state. Siderite, or any other susceptible material will respond to this field by producing a vectorless induced magnetization. This will be interpreted as a vector-based remanence magnetization by the superconducting coils. The susceptibility is unaffected by either AF or thermal demagnetization so that the signal observed by the pickup coils from the susceptibility will not vary from one measurement to the next. If susceptibility is substantially greater than the remanence, the cryogenic magnetometer will invariably return an apparent remanence magnetization direction that is constant with respect to the orientation marks on the specimen. Thus, the resultant apparent remanence direction and intensity is virtually identical every time the specimen is measured, leading to the false observation that the specimen retains a stable magnetization. The susceptibility effect is easily recognized because all specimens from a hand sample give the same direction, but each sample
gives a different direction. The data from one hand sample to another can be
“corrected” to a common direction using their declination and inclination of
orientation. This was the case for virtually all of the ore specimens in this
collection.

Several attempts were made to cancel out or account for the
susceptibility using multiple measurements in opposing orientations.
However, the susceptibility was so much greater than the remanence that even
a minor fluctuation in the measured value of the susceptibility introduced an
error large enough to overwhelm the underlying signal of the remanence
magnetization. Several specimens were measured using a Schonstedt
spinner magnetometer because the signal from this type of magnetometer is
unaffected by induced components. Unfortunately the remanence
magnetizations were not strong enough to be reliably measured on this
device.

The ChRM directions for the sites collected from the Cretaceous
Gem stock did crudely fit the Cretaceous APW path. However, since this
principle or ore portion of the project did not reveal any useful data, it was
decided not to proceed further. The remaining specimens in the collection
were not demagnetized and are still in storage. Should the sensitivity of
non-cryogenic magnetometers increase substantially or else if the problem of
the trapped fields in the cryogenic magnetometer can be overcome, it may be
possible to resurrect this project at a future date.
Appendix B

The Jerritt Canyon Project

A paleomagnetic study was attempted on the "Carlin Type" gold ore of the Jerritt Canyon mining district in the Independence mountains of northeastern Nevada. The purpose of this project was to attempt to extend the paleomagnetic dating method beyond MVT ore deposits. Unfortunately, owing to problems explained below, the results of this study were inconclusive.

The Jerritt Canyon gold mining district is located about 50 km north-northwest of the town of Elko in northeastern Nevada, U.S.A. (figure B.1) It is similar in style of mineralization and structural relationships to the ore deposits of the world-famous "Carlin Trend" gold ore district, located about 75 km to the southwest of the Jerritt Canyon district. Comparison of the Jerritt Canyon and the Carlin Trend districts shows that both are hosted by Ordovician to Silurian carbonates in a thrust sheet between the Golconda thrust to the west and the Robert's Mountain thrust to the east. The gold ore
Figure B.1

Map of northeast Nevada showing the location of the Jerritt Canyon gold district to the two major thrust faults. The stars show the locations of actively mined "Carlin Type" disseminated gold deposits in the region.
in both districts is found in disseminated tabular bodies with associated gangue mercury, zinc and arsenic. Thus, there is good reason to believe that a genetic relationship exists between these districts. These similarities can be extended to the other gold ore deposits in the region (figure B.1).

The age of these deposits has been debated. Researchers, working in both Nevada and Utah, have argued for ages ranging from as old as \( \sim 150 \) Ma (Presnel, 1992; Wilson and Parry, 1994), through \( \sim 117 \) Ma (Arehart et al., 1993), to as young as 34-38 Ma (Maher et al, 1993), despite the fact that, given the similarities from one deposit to another, it could be argued that they were all formed at the same time.

There have been several attempts to date the ore at Jerritt Canyon using radiometric methods, however, none have successfully dated either the ore stage minerals or their alteration products. Structural and alteration relationships of the ore to crosscutting dikes and sills that have been radiometrically dated, suggest that the ore is \( \leq 40 \) Ma at Jerritt Canyon (A. Hofstra, pers. com.) although the evidence is equivocal.

Paleomagnetic methods were employed in an attempt to date the gold ore. Four oriented hand samples were collected at each of 37 sites. Given the complex deformational history of the Independence mountains, the approach was to try to obtain ChRMs of \( \sim 40 \) Ma volcanic rocks on the eastern flank of the range (7 sites), the \( \sim 40 \) Ma dikes within the district (11 sites), and the ores (16 sites). These would then be compared to the North American
APW path to date the ChRMs and to identify any local fault block rotations, if present. At least three cores were obtained from each hand sample so that there were at least 12 specimens for each site for a total collection of over 500 specimens.

After measuring the NRM of the whole collection four pilot specimens from each site were selected for detailed demagnetization. Two pilot specimens were demagnetized in 12 steps up to a peak field of 150 mT. Where possible the other two pilot specimens were thermally demagnetized in 12 steps to 585°C. Unfortunately, the arsenic in the ore samples precluded their use in thermal demagnetization experiments because of the potentially harmful effects of the exhaust fumes. When thermal treatment was unavailable the remaining two specimens were AF step demagnetized as well. Upon analysis of these pilot specimens it was clear that there was virtually no magnetic material in the ore samples because their remanent magnetization was less than the noise level (~4 x 10^-6 A/m) of the magnetometer. When there was a measurable magnetization, it was invariably unstable both within a site and between sites. A few specimens did retain stable magnetizations but all were from barren rocks and they did not provide enough data to warrant further study on their own. It is speculated that the ore fluids had apparently replaced any preexisting magnetic minerals with nonmagnetic minerals. Although this was considered to be a possibility at the outset, it was surprising to find that there was no
measurable ChRM signature due to the gold hydrothermal system. The
paleomagnetic results are consistent with other evidence for sulphidization of
host-rock iron in these deposits (Hofstra et. al., 1991, Hofstra, 1994). The
results suggest that paleomagnetic methods are unlikely to be successful in
directly dating most, if not all, Carlin-type gold deposits. The ChRM
directions of pilot specimens from the volcanics and dikes, when barren, were
easily isolated and consistent with their known age. However, further work
on the remaining specimens from the volcanics and dikes was not thought
worthwhile because of the inability to obtain a ChRM for the ore.

The project was abandoned at this stage as there was little -r no
hope of determining the age of mineralization by paleomagnetism.
Appendix C

The "Geology" Article

The following four pages contain a reproduction of the article published in Geology in February of 1995. The article, co-authored with my thesis supervisor, represents the first attempt to apply Bingham (1974) statistics to paleomagnetic data to estimate the time required to form a typical MVT ore deposit. The conclusions presented in this article formed the basis for the majority of the research that comprises this thesis.

A mathematical error was found in the original article. The rate of APW should have been expressed as 2.7 Ma/° instead of 0.35 °/Ma, based on the reference poles of Van der Voo (1993). All data in this thesis use the correct values for the rate of APW. An errata for this article is in progress and should be published in Geology by the end of 1996.
Age and duration of Mississippi Valley–type ore-mineralizing events

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ABSTRACT

Statistics of a combined paleomagnetic data set from six studies of Mississippi Valley–type deposits and their surrounding host rocks show that the characteristic remanent magnetization of the host rocks predates the magnetization of the mineralization, providing a positive control test confirming that the ore magnetization is primary. Either simple fluid or dramatic changes to the fluid through time are required to account for the different magnetization ages observed in the host rocks and ores. The dispersion of the characteristic remanent magnetization directions provides an estimate for the duration of the mineralizing process, suggesting a mineralizing event of about 4 m.y. duration.

INTRODUCTION

The chemical and physical conditions during the formation of Mississippi Valley-type (MVT) Pb-Zn-Ba-F mineralization have been studied extensively (Everett, 1983; Cawthorn and Smith, 1983; Sangster, 1983; Leuch and Rowan, 1986), but the ages and durations of these events have remained equivocal (Ohle, 1989; Sangster, 1990). Recent paleomagnetic studies (Pan et al., 1990, 1993; Symons and Sangster, 1991, 1992; Pan and Symons, 1993) have determined mineralization ages but lack stringent conventional paleomagnetic stability tests. Additional information about the genesis of these ores may be obtained by analyzing several districts together and by applying Bingham (1974) statistics.

STATISTICAL METHODS

Dispersion within paleomagnetic data can be divided into random and nonrandom sources. Random dispersion is a function of sampling and measurement precision plus short-term (<10° m.y.) secular variation of Earth’s magnetic field, and it can be considered as the true error. Nonrandom dispersion can be caused by several factors, such as structural distortion, rifting, and apparent polar wander (APW), and it generates bias within a population. When Fisher (1953) statistics are used, it is assumed that population dispersion is caused solely by random error, whereas when Bingham (1974) statistics are used, it is assumed that both random and nonrandom variations occur in the population.

Traditionally, Fisher statistics have been used by paleomagnetists despite the fact that Bingham statistics provide a better description for many paleomagnetic populations. The reasons are that Fisher statistics can be easily computed using simple computer programs, their interpretation is straightforward, and the calculated confidence circles give conservative error estimates. Thus Bingham statistics have been avoided except where necessary. This study is an exception because Bingham statistics are robust in the presence of nonrandom influences and provide additional information that Fisher statistics cannot recognize.

When Fisher statistics are used, it is assumed that the unit vectors are randomly distributed about their mean with a Gaussian distribution away from the mean in all directions, to give a sphere of 95% confidence (radius = a 95) about the mean (Fig. 1). When Bingham statistics are used, it is recognized that the unit vectors may have a biased or elliptical distribution relative to the mean. Thus the 95% confidence interval is an ellipse with a semimajor axis of a 1-2 that also defines the bias direction, and a semiminor axis of a 2-1, that is perpendicular to a 1-2 (Fig. 1). When remanence directions are distributed along a linear trend, their skew will be reflected by the Bingham but not by the Fisher statistics. For the MVT deposits analyzed in this paper, the only plausible mechanism for their biased distributions is APW, because they occur in flat-lying undeformed sediments, thus precluding the possibility of a metamorphic or tectonic bias.

A simple estimate of the remanence acquisition time can be made by comparing the lengths of a 1-2 and a 2-1 axes because both estimate random dispersion but the major a 1-2 axis also includes dispersion from APW. The diameters of the ellipse are obtained by doubling a 1-2 and a 2-1. Subtracting 2a 1-2 from 2a 2-1 leaves an estimate of APW in degrees during remanence acquisition. If the rate of APW can be estimated, then the duration of remanence acquisition can be calculated.

ANALYSIS

Regional chemical remagnetization of sedimentary rocks has been shown to occur (McCabe et al., 1983) and has been attributed to...
to fluid flow caused by mountain building (Ozver, 1986). Thus the Kiaman remagnetization of sedimentary rocks (McCabe et al., 1983), coal maturation, and oil, gas, and MVT deposits on North America's interior craton are associated with the impact of the African and South American cratons with North America in Pennsylvanian-Permian time (Ozver, 1992). These features were caused by uplift and fluid expulsion from the formation of the Appalachian and Ouachita Mountains. Paleomagnetic pole ages coincided with nearby orogeny for both ore-stage minerals and their host rocks in several MVT districts throughout North America, but their remanences have been interpreted as pre-ore (Wilczewski et al., 1983), post-ore (Bashnoff et al., 1987), and most recently as the age of mineralization (Pam et al., 1990, 1992; Symons and Sangster, 1991, 1992; Pam and Symons, 1993). Sverjensky and Garven (1992) and others have argued for the relation between nearby orogeny and mineralization using information other than paleomagnetism.

Recent data were analyzed (Fig. 2, Table 1) from the Polaric-Northwest Territories (Symons and Sangster, 1992), Guys River-Nova Scotia (Pam et al., 1993), Newfoundland Zine (Pam and Symons, 1993), northern Arkansas-tri-state area (Kansas, Missouri, Oklahoma) (Pam et al., 1990), and central Missouri (Symons and Sangster, 1991) MVT districts, plus new data from the central Tennessee district, by applying Bingham (1974) statistics to site mean virtual geomagnetic poles (VGP's). Two additional MVT districts that have paleomagnetic results were excluded from this study. The east Tennessee study (Bashnoff et al., 1987) is in a structurally complex district, and the Pine Point (Symons et al., 1993) results were extremely weak and too scattered for Bingham (1974) analysis.

Comparison of the mean poles for the host rock and ore for each of the six ore districts in this way results in the recognition of three trends.

**OBSERVED TRENDS**

First, the mean host-rock pole invariably falls on a slightly older section of the North American APW path than the corresponding ore pole, although the 95% confidence limits overlap in three of the six districts (Fig. 2, Table 1). No rigorous tests are available for

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Table 1. Paleomagnetic Statistics

Note: Host rock: Oreo material; State is the stratigraphic age (Ordovician or Mississippian); N is the number of sites; a_p and a_s are the lengths of the poles of the 95% confidence; A_95 is the axis of the poles expressed in degrees from north. Central Tennessee results include six sites from a regional study of Tennessee carbonate (McCabe et al., 1989).

Bingham statistics to distinguish between two populations when the confidence intervals overlap. However, a simple empirical test of the data is whether the respective 95% confidence ellipses incorporate the other mean. By this test, two of the three districts with overlapping confidence intervals are distinct at the 95% level, and the third (central Missouri) plateau at the 63% level but is marginal at the 95% level. Therefore, because five of the six districts are distinct and the sixth is not definitive, these data show that the host-rock characteristic magnetization produces the ore magnetization.
Second, the magnetization of the ore and remagnetization of the host rocks appear to be temporally coupled in all districts except the northern Arkansas-tri-state district, despite much older depositional ages for the host rocks in all six studies, suggesting that the magnetization in the host rocks is genetically related to mineralization.

Third, the dispersion of the mineralization site mean VGP's about the unit mean has a preferred elongation direction that, except for Gays River, tends to parallel the APW path, indicating that the ore magnetizations were acquired over time sufficient to record APW.

Recognition that the ore characteristic magnetization differs from its host-rock magnetization provides a positive geochronological contact test (Everett and Clegg, 1962), confirming recent paleomagnetic studies (Pan et al., 1993, 1993; Symons and Sangster, 1991, 1992; Pan and Symons, 1993) which show that the ores retain primary magnetizations. This difference is in agreement with that recently observed for MVT mineralized veins intruding carbonates of the Viola Formation, Oklahoma (Elmore et al., 1993).

This difference further means that it is invalid to use host-rock magnetization directions to estimate the age of mineralization as was done in several studies (Symons and Sangster, 1991, 1992; Pan et al., 1993; Pan and Symons, 1993). The pole positions, confidence statistics, and age estimates given here supplant the previously published values.

DISCUSSION

The ore-magnetization ages for the six districts span >150 m.y., and these ores are emplaced in sediments that are as much as 200 m.y. older (Table 1), but within each district the host-ore pairs differ by only a few million to a few tens of millions of years. This suggests a genetic link between host-rock remagnetization and ore precipitation. Fluid-inclusion data and conodont alteration indices (Sangster et al., 1994) for the host rocks indicate mineralization or later temperatures that are far too low to thermally reset an existing remanence (Pulliahs et al., 1975). Thus, any remagnetization of the host rocks must have been the product of chemical interaction involving fluids. The distinct host-rock and ore characteristic remanent magnetization directions indicate that the fluid responsible for the host-rock remagnetization did not simultaneously precipitate the ore minerals. Either the fluid and/or its environment changed dramatically or a second fluid passed through each district. Thus, the second fluid or later phase of a single evolving fluid must be either incoercive with respect to the magnetite in the host rocks, or very localized, which conflicts with the regional distribution of trace MVT mineralization (Coveney et al., 1987). The lack of interaction between acidic ore-bearing brines and host carbonates can be explained by a high CO₂ content in the brines (Leach and Rowan, 1991).

The concept of multiple or evolving fluids has been suggested by several authors on the basis of geochemistry (e.g., Viets and Leach, 1990; Leach and Rowan, 1991). In an isotopic study of carbonates and MVT showings in Missouri, Blanner et al. (1988) argued that hot exhalational brines migrated to shallow depths, causing pervasive Mississippian dolomitization; later, carbonate and MVT mineral precipitation occurred in vugs and caves from passive brines without significant water-rock interaction. Cander et al. (1988), working in the same area, showed that although there are multiple generations of dolomitization, they all predate the MVT mineralization event in latest Pennsylvanian or earliest Permian time.

In most of the MVT districts, palaeomagnetic "breccia tests" (Graham, 1949) were attempted but failed; the clasts yielded a coherent magnetization direction. This shows that the host-rock remagnetizing fluid was active after the clasts were deposited, so that brecciation in these districts must predate or at least be contemporaneous with acquisition of the host-rock characteristic remanent magnetization.

The preferential elongation of the magnetization directions along the APW path, particularly for the ores, requires that both the host-rock remagnetization and ore-forming processes span at least a few million years in each district. Assuming an APW rate of 0.35°/m.y. for the later Paleozoic (Van der Voo, 1990), the ore-mineralizing process took from 1 to 8 m.y.; the average for the six studies (Table 1) is 4.4 ± 2.6 m.y. (root mean square error). This time frame agrees with that suggested by Garven et al. (1993), on the basis of hydrogeology.

This time calculation is not applicable to the host rocks because their characteristic remanent magnetization could have been updated continuously as the fluid-rock interaction proceeded. At a minimum, the preferential elongation of their magnetization directions requires that the remagnetization process take at least as long as mineralization.

Given the time interval that we postulate for mineralization and its acquisition of normal remanent magnetization, one would normally expect to record reversals of Earth's magnetic field in these studies. The Kiaman superchron was a prolonged period of almost exclusively reversed polarity from mid-Carboniferous (~320 Ma) to latest Permian (~250 Ma). Its boundaries are not tightly constrained. Irving and Pulliahs (1976) placed the bounds at 313 and 227 Ma, and Cox (1982) suggested 320 and 250 Ma. The compilation of global paleomagnetic data by Van der Voo (1993) listed only 16 of 122 studies from stable cratonic settings between 320 and 250 Ma that have some specimens or sites with normal polarities preserved in them. Northern Arkansas, central Missouri, and Gays River mineralization poles fall fully within the Kiaman, so only the reversed polarities would be expected. The mineralization age in central Tennessee straddles the end of Kiaman time. Although most of its ore specimens record a reversed polarity, several do retain the opposing normal polarity. The normally magnetized Newfoundladian Zois reverse magnetized Polaris ores indicate Devonian ages, although neither shows reversals. Although both polarities have been recorded in Devonian rocks, there is no well-established magnetostatigraphic record. Thus, the absence of reversals within these two deposits is not good evidence for a shorter mineralizing event.

SUMMARY

Our analysis shows that MVT ores retain primary magnetizations that were acquired during a localized mineralizing event lasting from 1 to 8 m.y., and that mineralization occurred after the completion of pervasive regional host-rock remagnetization. We note that the temporal relation of the host-rock and ore characteristic remanent magnetization has been observed over wide areas of North America in five carbonate-platform-organogenic belt pairs: the Ozark dome and Ouachita, the Nashville dome and Appalachia, the Subarcsides basin and Appalachia, the Humber zone and Appalachia, and the Sverdrup basin and Inuvikia. Thus, the fluid-rock relation observed in Missouri (Blanner et al., 1988; Cander et al., 1988) between earlier nonmineralizing fluids which were followed by or evolved into later mineralizing fluids appears to be characteristic of many MVT mineralized platforms.

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Appendix D

Hysteresis Properties of MVT Minerals and Carbonate Rocks

In recent years it has become clear that a vast portion of the Paleozoic carbonates in the midcontinent of North America were remagnetized during the Kiaman reversed superchron, a period of almost-exclusively reversed magnetization that lasted from Mississippian into Permian time. Although the existence of the remagnetization has now been accepted, the mechanism for remagnetization and the characteristics of the magnetic carriers have remained controversial. This has led to a number of paleomagnetic and rock magnetic studies that have attempted to better identify remagnetized carbonates and to understand the processes that led to their remagnetization.

Jackson (1990) measured the hysteresis properties and the ratios of coercive force to remanence coercivity (Hc/Hcr) and of saturation remanence
to saturation magnetization (Mr/Ms) for remagnetized Paleozoic carbonate rocks from the Trenton, Onondaga, and Knox formations of the midcontinent of the U.S.A.. He found that these samples had anomalously high Hc/Hcr and Mr/Ms ratios as well as unusual constricted or "wasp-waisted" hysteresis loops. Constricted hysteresis loops are characterized by a narrowing or necking in the middle section of the loop near the origin with a wider separation both above and below the origin until they close as saturation is approached. He argued that these properties were the product of fine grained, single domain (SD) magnetite coexisting with either extremely fine superparamagnetic (SP) magnetite or coarse-grained multidomain (MD) magnetite. A mixture of SP and SD magnetite can be obtained from a single generation of magnetite with a size distribution that spans the SP-SD threshold (~ 0.05 µm). However a SD-MD mixture would likely require two generations of magnetite with differing size distributions and, therefore, two distinct magnetite-forming events as one might expect to find in chemically remagnetized carbonates. Thus, if the constricted hysteresis loops were the product of a SD-MD mixture, then it could be diagnostic of a chemically remagnetized rock. Initially Jackson (1990) suggested that a SD-MD mixture was present. Later, Jackson et al. (1992) used the ratio of anhysteretic remanence to saturation remanence to infer the existence of a population of magnetite that spans SP-SD threshold in addition to a SD-MD mixture.
McCabe and Channell (1994) and Channell and McCabe (1994) compared carbonate rocks that are known to be remagnetized with those that retained their depositional remanence magnetizations. They showed that the diagnostic shape of hysteresis loops, first recognized by Jackson (1990), was present in remagnetized carbonates from Britain, Alaska and Nevada. More importantly, they found that this behaviour was absent in limestones from Italy that carry a primary magnetization. Their results confirmed on a wider scale that constricted hysteresis loops are diagnostic of remagnetized carbonates.

Recently Roberts et al. (1995) showed for rocks with hysteresis loops that are closed (i.e. completely saturated) by 0.3 T as has been the case for most carbonates, that the “wasp-waisted” characteristic can be attributed to the existence of two phases of magnetite rather than a magnetite-hematite combination.

The experiment described here intended to compare the hysteresis behaviour of fine-grained dolostones, sparry or hydrothermal dolomites, and pure crystalline forms of MVT mineralization. If the properties of fine-grained dolostones that are known to be remagnetized, differed from those of hydrothermal dolomites and MVT ore crystals then it would provide further confirmation that the ore-stage material retained primary remanences.
Several attempts were made to obtain hysteresis loops for the CT suite of rocks. Measurements were made on whole-rock samples only. Magnetic separates were avoided because of concerns that the finer fractions might be disproportionately lost during the separation process. First, several 1.2 cm diameter by 1.0 cm height right-cylindrical specimens were measured on a custom-built vibrating sample magnetometer (VSM) at the Institute de Physique du Globe in Paris, France. Second, the samples were measured at the Institute for Rock Magnetism (IRM) in Minneapolis, Minnesota, using both a Princeton Measurements Corporation VSM on 2.5 cm diameter by 2 cm height specimens and on a Princeton Measurements Corporation alternating gradient force magnetometer (MicroMag) on small (<100 mg) rock chips. Finally, an additional set of measurements were made at the Erindale College lab of the University of Toronto, Ontario, also using a MicroMag on rock chips.

The initial experiments in Paris proved fruitless. The equipment was not sensitive enough at the time to generate reliable data for any of the rock types. Several specimens were left behind to use as test samples as the equipment was upgraded. The experiments conducted at the IRM proved that there was virtually no magnetic material in the pure crystalline specimens and that the dolomites and dolostones were barely above the noise level of both the VSM and the MicroMag. The final experiment, conducted at the University of Toronto confirmed the results from the first two labs.
Figure D.1

Sample hysteresis loops showing: (a) an empty sample holder; and, (b) a sample from a single crystal of sphalerite.
Figure D.1a shows the hysteresis loop for a typical run on an empty sample holder from the Toronto lab, after subtraction of the diamagnetic/paramagnetic contributions. Since it was known in advance that the specimens were very weakly magnetized, frequent measurements of the holder were made and these were subtracted from subsequent measurements of the samples. Notice that the background values for Mr and Ms are ~60 pAm$^2$ and 600 pAm$^2$ respectively. All subsequent measurements shown here were corrected for the contribution of the sample holder.

Example loops for pure crystals of sphalerite (figure D.1b), fluorite (figure D.2a), and barite (figure D.2b) show clearly that the crystalline form of MVT mineralization in the Elmwood-Gordonsville deposit does not contain any magnetic material so that further work on these samples was useless.

An example of one of the most intensely magnetized hydrothermal dolomites (figure D.3a) shows that, while slightly better results were obtained, the signal to noise ratio was still far too low to identify any shape characteristics in the hysteresis loop or place any confidence in the Mr/Ms or Hc/Hcr ratios.

Finally, three examples of fine-grained dolostone are shown with one each from the Toronto (figure D.3b), Paris (figure D.4a) and IRM (figure D.4b) labs. The Paris data was obtained by B. Henry and M. LeGoffe after their VSM was updated and improved. All three of these samples yielded similar results. Together, they suggest the possibility of a constricted
Sample hysteresis loops showing: (a) a sample of crystalline fluorite; and, (b) a sample of crystalline barite.
Figure D.3

Sample hysteresis loops showing: (a) a sample of hydrothermal dolomite; and, (b) a sample of fine grained dolostone.
Sample hysteresis loops showing two samples of fine grained dolostones. Notice that the appearance of a constricted hysteresis loop is suggested in b.
hysteresis loop. The sample from the IRM is the most intensely magnetized and provides the best example of the three; however, even for it the data were too poor to infer ratio characteristics that might separate it from rocks that have not been remagnetized. It is interesting to note that of all the previously published data discussed here, Jackson’s (1990) data from the Knox supergroup was the poorest in terms of resolution of the shape of the hysteresis curve as well as giving the weakest saturation magnetization.

The data presented here are consistent with previous trends for remagnetized carbonates although they are not sufficient to compare the unmineralized and mineralized samples from the central Tennessee area. However, it can be concluded that the pure crystalline samples of MVT mineralization do not contain sufficient magnetic minerals in their structure to give any response using the best instrumentation that is currently available.
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