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DEPARTMENT OF MINES
AND TECHNICAL SURVEYS

BULLETIN 71

# PALAEOMAGNETISM AND CORRELATION OF KEWEENAWAN ROCKS

P. M. DuBois

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1962

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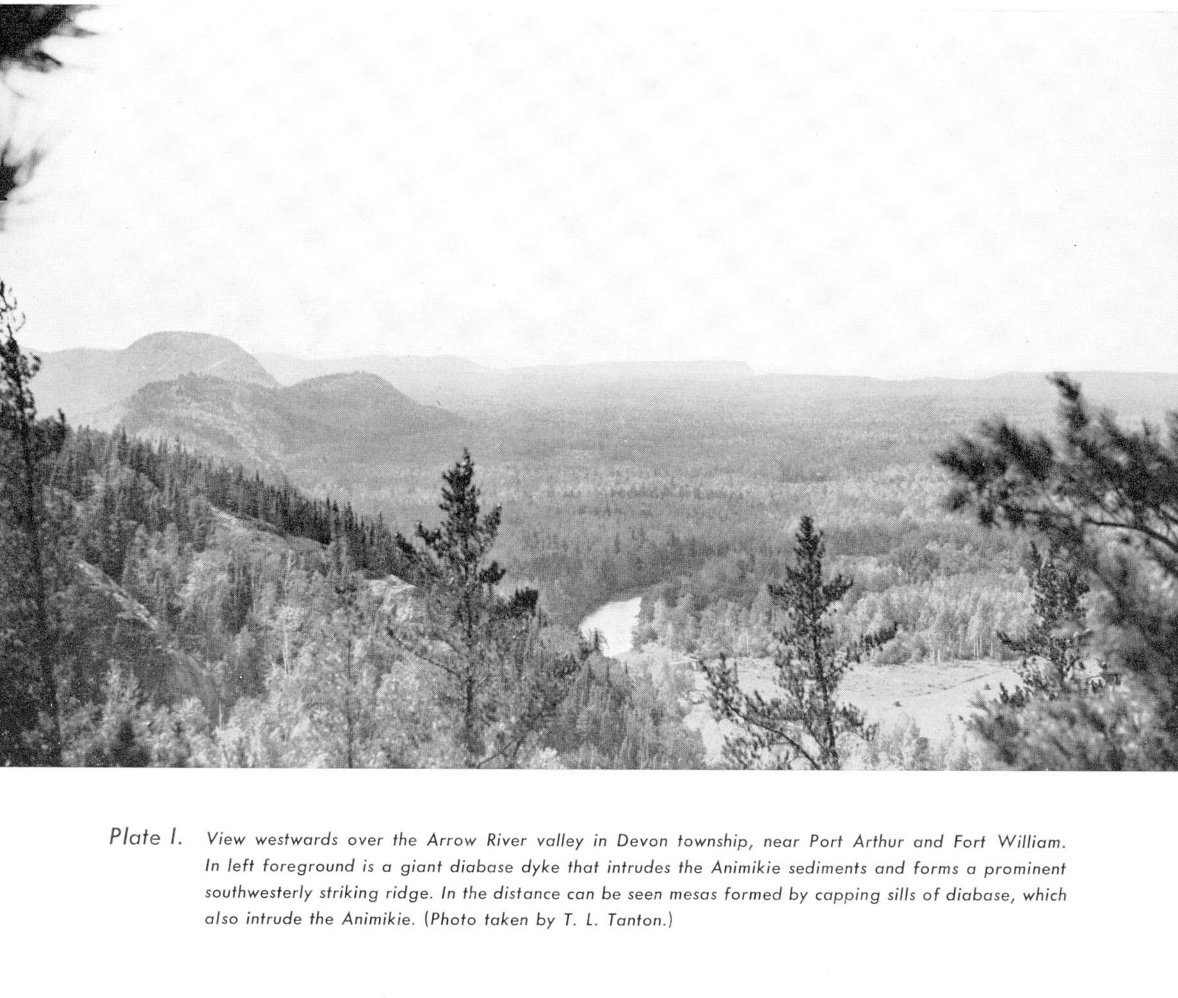

Plate I. *View westwards over the Arrow River valley in Devon township, near Port Arthur and Fort William. In left foreground is a giant diabase dyke that intrudes the Animikie sediments and forms a prominent southwesterly striking ridge. In the distance can be seen mesas formed by capping sills of diabase, which also intrude the Animikie. (Photo taken by T. L. Tanton.)*

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# PALAEOMAGNETISM AND CORRELATION OF KEWEENAWAN ROCKS

By
P. M. DuBois

DEPARTMENT OF
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# PREFACE

One of the most difficult problems and perhaps the most important of the geologist working in Precambrian terrain is the chronological correlation of rock-units from widely separated localities. Until recently this could rarely be done, although it was frequently attempted on the misleading basis of lithological similarity. The development of age determination by isotopic ratio marked the first major advance in correlating Precambrian formations. Another promising and independent method, recently developed, is based on determining the position of the earth's magnetic pole at the time the rock-units were formed.

The work described in this bulletin is an attempt to effect a correlation, based on such palaeomagnetic studies.

J. M. HARRISON,
*Director, Geological Survey of Canada*

OTTAWA, May 11, 1960

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# PALAEOMAGNETISM AND CORRELATION OF KEWEENAWAN ROCKS

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## *Abstract*

The directions of the remanent magnetism of pre-Tertiary rocks are different from the direction to be expected from an axial geocentric dipole with the present axis of rotation of the earth. This difference is interpreted as a movement of the geographic and geomagnetic poles relative to the collecting sites. After the path followed by the pole during a geological period has been accurately determined, it should be possible to discover the relative age of a geological unit formed during that period by determining, from the remanent magnetism, the position of the pole when the rocks were formed.

In this study 400 oriented specimens were collected from one of the thickest and best known successions of the Precambrian rocks in the Lake Superior region. The palaeomagnetic directions of some of these specimens served to trace the curve of polar wandering during that part of the Precambrian known as the Keweenawan. The relative ages of other Keweenawan samples collected in areas where the age and correlation were in doubt, were determined by comparing their direction of magnetization with the polar wandering curve established. The study was extended to include palaeomagnetic measurements of Precambrian specimens from the Boulter and Umfraville-Thanet gabbros of the Bancroft area, the norite of the Sudbury basin, the Belt Series of Alberta, members of the Grenville series, the Hakatai shale of Arizona, and the Hazel formation of Texas. The ages deduced from the palaeomagnetism are in complete accord with the known geological facts and the isotopic age determinations made for these or closely related rocks.

## *Résumé*

Les directions du magnétisme rémanent des roches prétertiaires diffèrent de celle que prendrait une roche, si elle était aimantée dans le dipole géocentrique actuel de la terre. Cette différence a été interprétée comme indiquant un déplacement graduel des pôles géographiques et magnétiques de la terre par rapport aux endroits où l'on a recueilli des spécimens. On devrait pouvoir, après détermination exacte de l'itinéraire suivi par le pôle au cours d'une période géologique donnée, déterminer l'âge relatif d'une unité géologique qui s'est formée pendant cette période, en déterminant, d'après le magnétisme rémanent, la position du pôle lors de la formation des roches.

Pour cette étude, 400 spécimens orientés furent recueillis de l'une des successions les plus épaisses et les mieux connues des successions précambriennes de la région du lac Supérieur. Les directions du paléomagnétisme de certains de ces spécimens ont servi à tracer la courbe de la migration du pôle au cours du Keweenawien qui est une subdivision du Précambrien. L'âge relatif d'autres spécimens de roches keweenawiennes provenant de régions où l'âge et la corrélation étaient douteux, fut déterminé en comparant la direction de leur paléomagnétisme avec la courbe mentionnée plus haut. En outre, on a mesuré le paléomagnétisme

de spécimens précambriens provenant des gabbros Boulter et Umfraville-Thanet de la région de Bancroft, de la norite du bassin minier de Sudbury, des couches Belt de l'Alberta, de niveaux des couches Grenville, du schiste Hakatai d'Arizona et de la formation Hazel de Texas. Les âges déduits de l'étude du paléomagétisme concordent parfaitement avec les événements géologiques connus et avec les radiodatations obtenues dans le cas de ces roches ou d'autres qui leur sont intimement apparentées.

## *Chapter I*

# INTRODUCTION

Palaeomagnetism may prove invaluable to the geologist as a tool for the correlation of non-fossiliferous formations, especially those of Precambrian age. Even after the appearance of the first useful fossil-forming organism in the Cambrian period, many non-fossiliferous rocks are difficult to date. Furthermore, most radioactive age determinations, except for radio-carbon and some potassium-argon dating, have been done on intrusive igneous rocks, and it is often difficult to place these rocks in their proper place in the geological column. In the Precambrian, the challenge is almost unlimited as there are no fossils with which to correlate sedimentary rocks that, if they were younger, would have been fossiliferous. Most Precambrian rocks have been correlated by the study and comparison of facies, unconformities, and rank of metamorphism. Correlations based on such evidence are very dubious. Thus, on the Canadian Shield, the flat-lying, relatively unmetamorphosed rocks have mostly been grouped together into the Proterozoic, which is by definition younger than the Archæan, into which the strongly folded and metamorphosed rocks are classified. However, in this scheme of classification some Archæan-type rocks may be younger than some Proterozoic-type rocks.

During recent years, much progress has been made in the measurement and interpretation of the natural remanent magnetization of rocks. It has long been known (Chevallier, 1925)[1] that rocks can acquire a stable magnetization in the direction of the earth's magnetic field. Thus, it is possible, by collecting oriented samples of rocks and measuring their directions of magnetization with very sensitive astatic or rock-generator magnetometers, to determine the direction of the earth's magnetic field at the time the rock was formed. The orientations of the magnetizations are randomly distributed and as the process of orienting, collecting, preparing, and measuring a sample introduces a random error of about 5 degrees, it is desirable to collect many samples from a geological formation to determine its mean direction of magnetization. By sampling many geological formations, it has been possible to determine the direction of the earth's magnetic field as a function of time even as far back in earth history as the Precambrian.

Measurements in late Tertiary rocks throughout the world show that there is a close correspondence between the earth's magnetic field averaged over a period of several tens of thousands of years and the field produced by an axial geocentric dipole (Hospers, 1953; Campbell and Runcorn, 1956; Irving and

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[1] Names and dates in parentheses are those of references listed in the Bibliography.

Green, 1957; DuBois, 1959). They also show that the earth's magnetic field has reversed its sign of polarity many times during the past. The present earth's magnetic field has in addition to its dipole component, which accounts for most of the field, quadrapole and other multi-pole components. These multi-pole components, which cause the earth's dip poles to differ from the geographic poles, can be explained by the fields produced by centres of secular variation, which are randomly distributed at the surface of the earth's core. The fields produced by these centres of secular variation, according to the late Tertiary results, average out to zero over a sufficiently long period of time leaving the geocentric axial dipole field as the mean field.

When the directions of remanent magnetization of pre-Tertiary rocks are examined, they are found to be different from that to be expected from an axial geocentric dipole with the axis of rotation in its present position. If the average magnetic poles and the geographic poles have coincided during all geological history as they apparently did during late Tertiary time, this difference must be interpreted as a movement of the geographic and geomagnetic poles relative to the place where the palaeomagnetic measurements were made. If this polar movement is constant in rate, a certain shift in the position of the pole will correspond to a certain interval of time.

The main problem is to determine accurately the path followed by the pole during geological time. The shape of the path can be determined simply by making sufficient measurements on rocks whose relative ages are known. To construct an absolute chronology, however, it is necessary to locate on this path enough points representing the position of the pole at known times so that reliable interpolations may be made between them. Cooperation between workers in radioactive dating and palaeomagnetism may thus enable rapid progress to be made in geochronology. This method of correlation can be used between two points far apart, as long as relative movements have not taken place between the two points. Moreover, correlations may still be made if the nature of these movements is known and corrections can be made.

In this paper an attempt will be made to show how palaeomagnetism can aid and supplement the geological interpretation of the Lake Superior region where one of the thickest and best known successions of Precambrian rocks is exposed. The work commenced with a field trip for collecting rocks during the summer of 1954, while the author was a research student at the University of Cambridge, and was finished during the summer of 1958, when the last samples, which were collected that spring, were measured at the Geological Survey of Canada. More than 400 samples were measured. The purpose of this investigation was two-fold: to see whether palaeomagnetic methods could be used to solve local geological problems and to start building a standard polar-wandering curve that could be used in future correlations of Precambrian rocks from other regions by palaeomagnetic techniques.

The paper is divided into three main parts: a brief description of the techniques of collecting, measuring and 'magnetic washing' of rock samples; a fairly lengthy and detailed account of the geological setting and the palaeomagnetic results; and a discussion of the significance of these results.

The author wishes to thank Professor S. K. Runcorn of the University of Cambridge, and many of his colleagues at both the University of Cambridge and the Geological Survey of Canada. Particular thanks are extended to R. F. Black, who did much of the laboratory work and who accompanied the author on one field trip to Lake Superior.

# *Chapter II*

# TECHNIQUES

## Collection and Measurement of Samples

More than 400 samples of rock were collected from the Keweenawan series and prepared in the following manner. The strike and dip of a plane surface, either a joint or a bedding plane, were marked on the surface, generally with a china-marking crayon. The azimuth of the strike line was then determined, generally with a Brunton compass although with fairly magnetic rocks, such as lava flows, or diabase intrusions, it was necessary to use a solar compass. The angle of dip was measured with the clinometer of a Brunton compass. At Cambridge the rocks were cored with a 1⅜-inch, non-magnetic, diamond trepanning tool, the core being taken perpendicular to the reference plane on which the strike and dip were marked. These cores were then cut into one-quarter-inch disks if the rocks were sedimentary and 1⅜-inch long cylinders if they were volcanic rocks. At the Geological Survey of Canada each rock was cut into one-inch cubes with a non-magnetic circular diamond saw so that the top face was parallel with the reference plane. If the rock samples were big enough, at least two specimens were cut. At Cambridge sometimes as many as a dozen disks were cut and measured.

At Cambridge the sediments were measured on a sensitive astatic magnetometer of the kind developed by Blackett (1952). A description of this instrument and the technique of measurement is given in a paper by Collinson, Creer, Irving, and Runcorn (1957), who also described a simple short period astatic magnetometer of lesser sensitivity, used to measure the volcanic rocks.

The measurements at the Geological Survey of Canada were also made on two different magnetometers. One was a sensitive Blackett type, astatic magnetometer developed and built for the Dominion Observatory by J. Roy and illustrated in Plate III; the other was a spinner or rock generator magnetometer (*see* Pl. II), the sensitivity of which was considerably less than that of the astatic magnetometer.

The technique of measurement on the Dominion Observatory astatic magnetometer differed somewhat from that described by Collinson, *et al.* Instead of offsetting the specimen to the east and west of the axis of the magnet system to determine the vertical remanent moment, the specimen was always kept on the axis of the magnetometer system and rotated about that axis. Furthermore, whereas Collinson, *et al.,* first measured the deflection of the magnetometer when the specimen was far away from the magnet system, then brought it close to the lower magnet, then removed it again, rotated it through 90 degrees and repeated

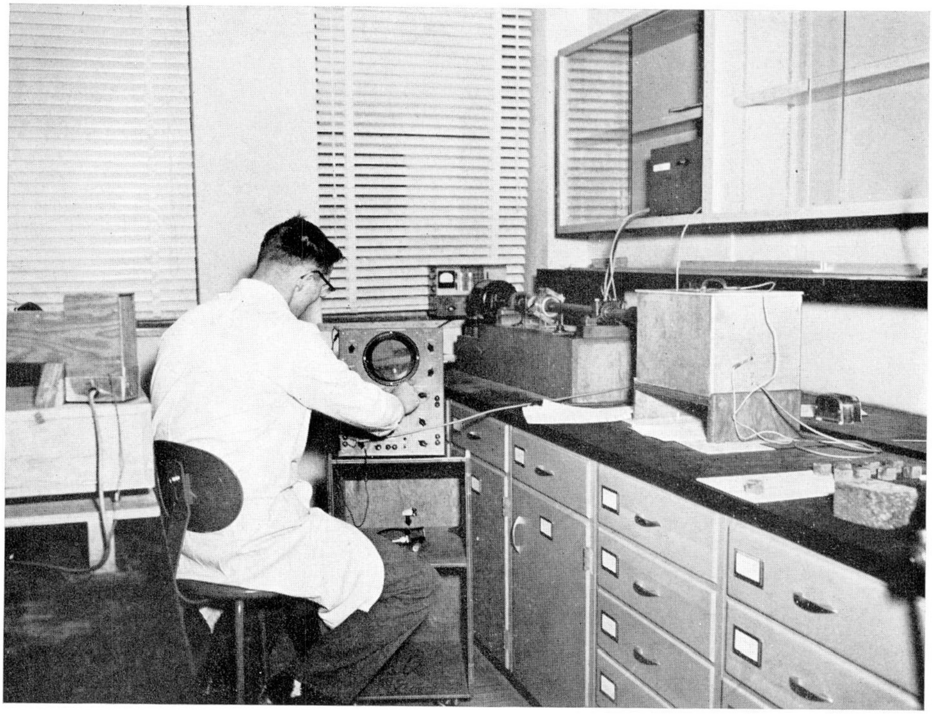

Plate II. *Spinner-type magnetometer.*

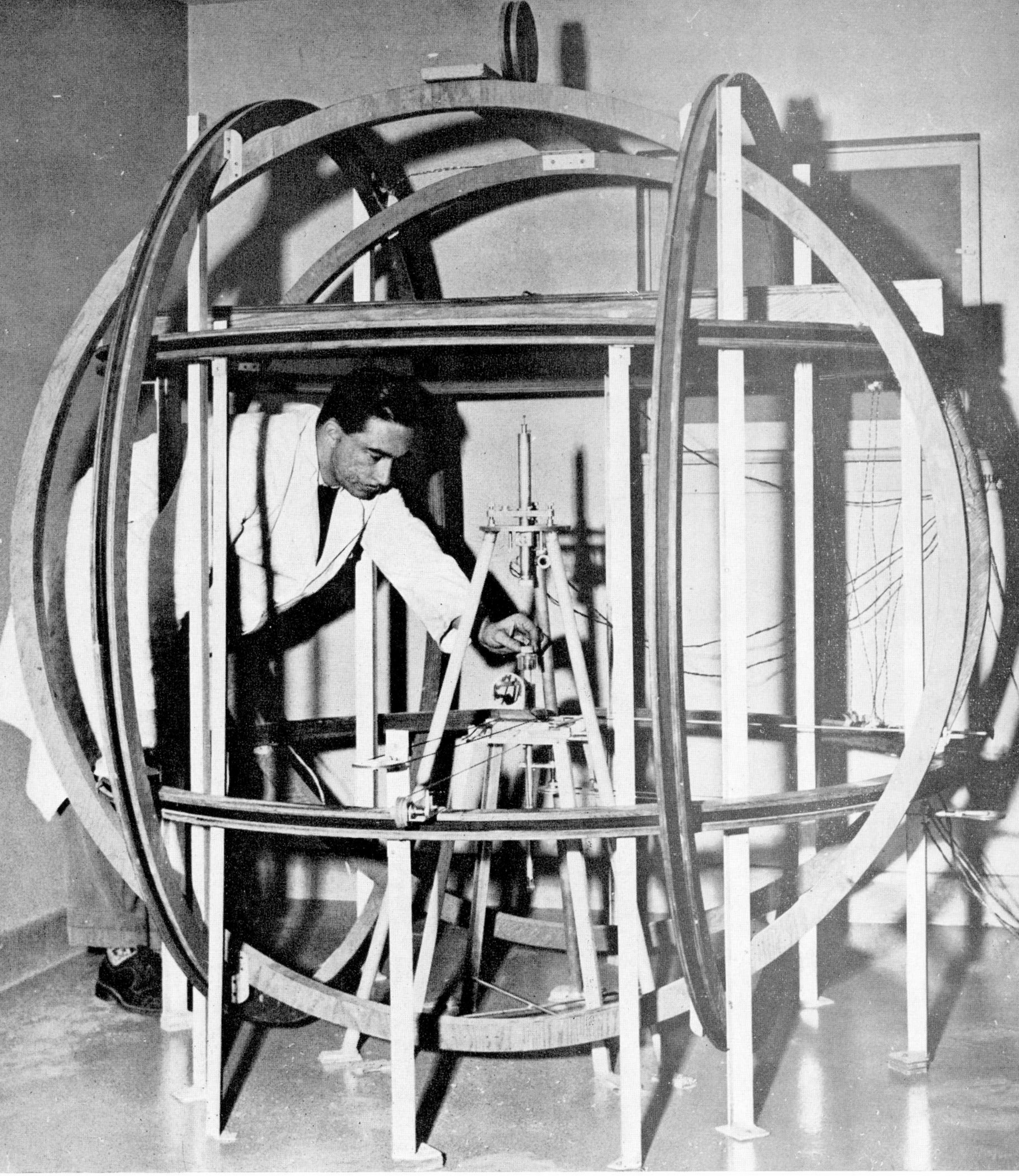

Plate III. *Astatic magnetometer. (Courtesy of Dominion Observatory, Ottawa).*

the process until 360 degrees was reached, all measurements on the Dominion astatic were made at constant height. Let $I_x$ be the remanent magnetic component along the reference line (which is parallel to the original strike line of the rock sample and to one edge of the cube) and $I_y$ be the component perpendicular to $I_x$ in a clockwise direction and in the original reference plane, which is now the top of cube (*see* Fig. 1). The cubical specimen is then placed under the magnet system, the axis of which is perpendicular to the reference plane and passes through the centre of the cube, so that the reference line is parallel with the lower magnet. A reading, $D_{0°}$, is then taken, and the specimen rotated through 180 degrees. A second reading, $D_{180°}$ is taken, and the specimen is rotated back through 180 degrees in which position a second reading at 0° is made, $D'_{0°}$. If the drift of the instrument is constant and as $I_x$, being parallel with the lower magnet, causes no deflection, we can write

$$2I_y = C\left(D_{180^\circ} - \frac{D_{0^\circ} + D'_{0^\circ}}{2}\right)$$

where C is a constant inversely proportional to the sensitivity of the instrument and to the cube of the distance from the lower magnet. C may be either positive or negative, the sign depending on the geometrical relationships of the magnetometer's optical system and the polarity of the lower magnet. A similar set of measurements can be made at 90 degrees and 270 degrees, and $I_x$ can be computed according to the formula

$$2I_x = C\left(D_{270^\circ} - \frac{D_{90^\circ} + D'_{90^\circ}}{2}\right)$$

The cube can then be rotated so that each of its six faces are placed directly under the magnet system, and for each position a similar set of measurements can be made. In this way four determinations of each of the three components $I_x$, $I_y$ and $I_z$ can be made.

The advantages of this technique are as follows:

(1) All measurements are made with the cube symmetrically placed with respect to the axis of the magnet system. Therefore, as the cube is rotated about this axis, the deflection of the magnetometer is a sinusoidal function of the angle of rotation. Such is not always the case with non-uniformly magnetized specimens of rocks measured by the method described by Collinson, *et al.* (1957), in which the rock is rotated about an axis that is displaced to one side or the other of the magnet system (Irving, 1954). As calculation of $I_x$ and $I_y$ depends on the assumption of such a sinusoidal function, any departure will introduce error, although some of these errors can be eliminated by averaging.

(2) The constant C, although it depends inversely on the third power of the distance from the centre of the cube to the lower magnet, is the same for all three components, if the rock is uniformly magnetized. Therefore, in computing the mean direction of magnetization it is not necessary to know C or the distance to the magnet system. However, in the method of Collinson, *et al.*, the distance from the magnet system to the specimen and also the displacement of the axis of

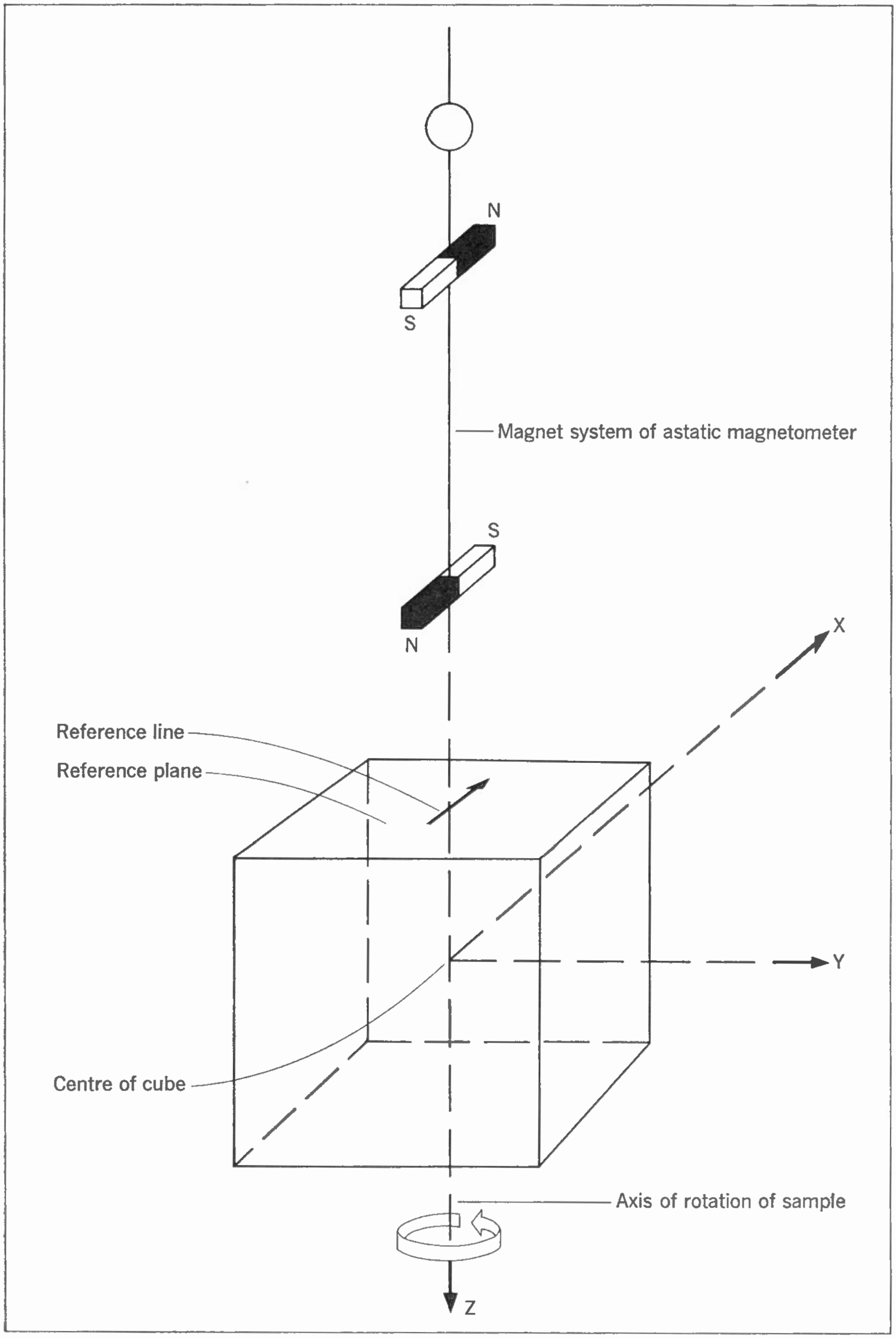

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*Figure 1. Arrangement for measuring $I_x$ and $I_y$.*

rotation of the specimen from that of the magnet system must be known explicitly in order to compute the vertical component. Any systematic errors in these quantities will therefore introduce systematic errors in the vertical component or angle of inclination.

(3) As the spinner magnetometer at the Geological Survey of Canada takes one-inch cubes, the use of such cubes in astatic measurements makes the two instruments interchangeable.

The spinner or rock generator magnetometer at the Geological Survey of Canada consists of a wooden specimen holder connected to one end of a long brass rod, which is spun by an electric motor at about thirty revolutions per second. Near the motor end of the brass rod a reference magnet is embedded in the rod with its axis perpendicular to that of the rod. A signal is picked up from the reference magnet by a coil system, whose axis is perpendicular to the brass rod and which can be rotated about the rod.

The signal generated by the rotating rock sample is picked up by a shielded pancake coil and fed to a high gain amplifier. The outputs of this amplifier and the amplified reference signal are fed to an electronic switch, whose output is fed to the vertical plates of a single beam oscilloscope. It is thus possible to compare the phase and amplitude of the sample signal with those of the reference signal. The two signals on the oscilloscope can be adjusted to have the same amplitude by varying the amplification of the high gain amplifier and attenuating the reference signal. They can then be brought into phase by rotating the reference pick up coil about the brass rod.

The measurements are made by spinning the cube about three orthogonal axes perpendicular to the faces of the cube and recording the phase angle and amplitude of the signal. From these two quantities two of the three components, $I_x$, $I_y$, and $I_z$, can be computed, and for each component two values are obtained which can be averaged. This magnetometer can measure rock samples with intensities of magnetization of $10^{-4}$ emu. $cm^{-3}$, whereas measurements with the astatic magnetometer can be made on intensities as low as $10^{-6}$ emu. $cm^{-3}$. A number of comparative measurements were made on cubes with both types of magnetometer and no significant differences were found in the measured directions.

All the data reported in this paper are presented in terms of declination, D, measured in degrees east of geographic north and inclination, I, which is positive in the downward direction and negative in the upward. All directions are corrected for geological dip where necessary unless otherwise stated. The circles of confidence, X, have been computed by Fisher's statistics at the 95 per cent level of confidence, and, in the various tables, K and N refer to the precision and the number of samples respectively. In plotting the magnetic results graphically both hemispheres of a stereographic projection have been used. Full circles indicate north magnetic poles on the lower hemisphere and open circles represent similar poles on the upper hemisphere. The mean directions are also plotted with crosses for north poles on the lower hemisphere and open circles with rays on the upper hemisphere. The circles of confidence are drawn with solid and dashed lines for the lower and upper hemispheres respectively. The field direction produced at the locality by an axial geocentric dipole is represented by a solid triangle on the lower hemisphere.

## Magnetic Washing

Many samples of rock collected for palaeomagnetic research have remanent magnetic moments that do not form a consistent pattern when first measured. This inconsistency can commonly be ascribed to the combination of an original stable magnetization, acquired when the rock was first formed, and a secondary magnetization that may or may not be stable over a period of many years. Probably the most common secondary component is the result of an isothermal remanent magnetization induced by the earth's magnetic field in the rock sample over a period of time. If the relaxation time of this isothermal remanent magnetization is fairly long compared with the time between the collection of the rock and its measurement, normally a few months, then the induced isothermal remanence can be considered semi-stable, and the measured natural remanence will lie between the direction of the original stable magnetization and the average direction of the earth's magnetic field during recent geological time. If the proportion of original remanence to isothermal remanence varies from sample to sample, the measured directions will be distributed in a plane containing the two components. Sometimes the time of decay of the isothermal remanence is less than or equal to the length of time before the sample is measured. In this case, if the rocks are randomly stored after collection there will be superimposed on the original stable component a random distribution, which will not alter the mean direction for many samples but which will increase the circle of confidence. Sometimes, however, the original stable magnetization may be so much weaker than the induced component that it cannot be detected through the 'noise' of the random component.

Secondary components of magnetization may be superimposed on rock samples by various other mechanisms, such as lightning, chemical alteration, and reheating. Any component due to lightning is likely to be random, whereas components due to the last two mechanisms are likely to be consistent. In the cases of chemical alteration and reheating the acquired magnetization will be in the direction of the earth's magnetic field at the time of these events. It is now evident that the earth's magnetic field has changed during geological time. Therefore, if the chemical alteration and reheating took place in several episodes over a very long period of time, the superimposed components due to these mechanisms will become increasingly complex and in general more nearly random.

Thus, in principle, any rock may have a magnetization that is the vector sum of its original magnetization and one or more components, due to the above-mentioned mechanisms. Two methods have been used to eliminate these secondary components, namely thermal and alternating current demagnetization.

## Thermal Demagnetization

In thermal demagnetization a sample is heated by stages up towards its Curie point in field free space. Its direction and intensity of magnetization can then be measured either while it is being heated or after it has been cooled in a zero magnetic field and removed from the furnace. Although the latter process is more time consuming, it is technically simpler and does give certain information about the decay in the intensity of remanence that the first method does not. If an isothermal remanence has been superimposed on a rock sample, thermal demagnetization should begin to destroy it before decay of the stable natural remanence sets in. Roquet (1954) found that isothermal remanence will decay with a slight rise in temperature, whereas stable natural remanence or thermoremanence will not really begin to decay until the temperature nears the Curie point. A number of samples of Keweenawan Freda sandstone and Portage Lake lavas, whose magnetic stability has been shown by field tests, were heated and cooled in field free space. The decay of natural remanence with temperature is shown in Figure 2 in terms of the quotient of its intensity after heating to temperature T and cooling back to room temperature (20°C) by the original intensity. These results show that each rock sample has two ferromagnetic constituents of different Curie points and also show the typical step-like nature of the Curie decay of stable natural

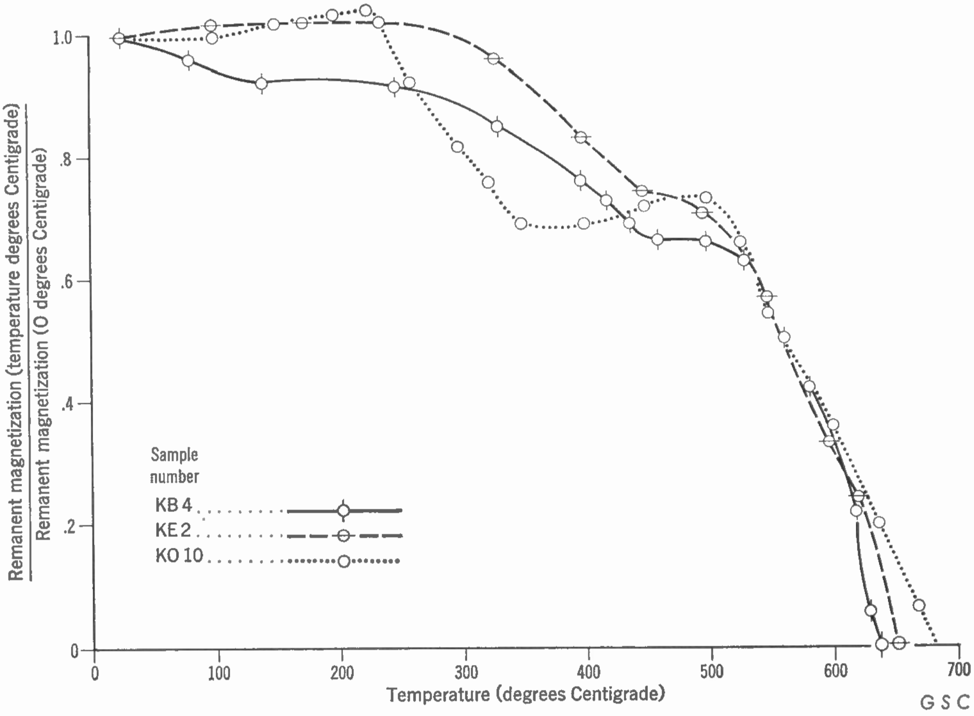


*Figure 2. Thermal demagnetization of Keweenawan sandstones and lava.*

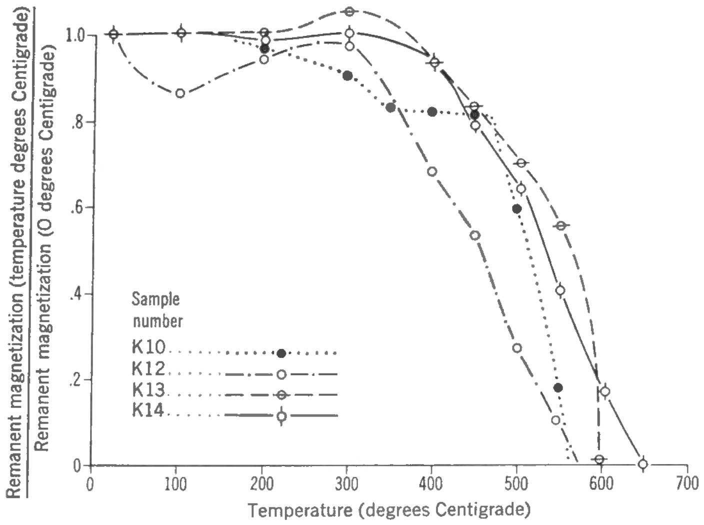


**Figure 3a.** *Thermal demagnetization of Chequamegon sandstone.*

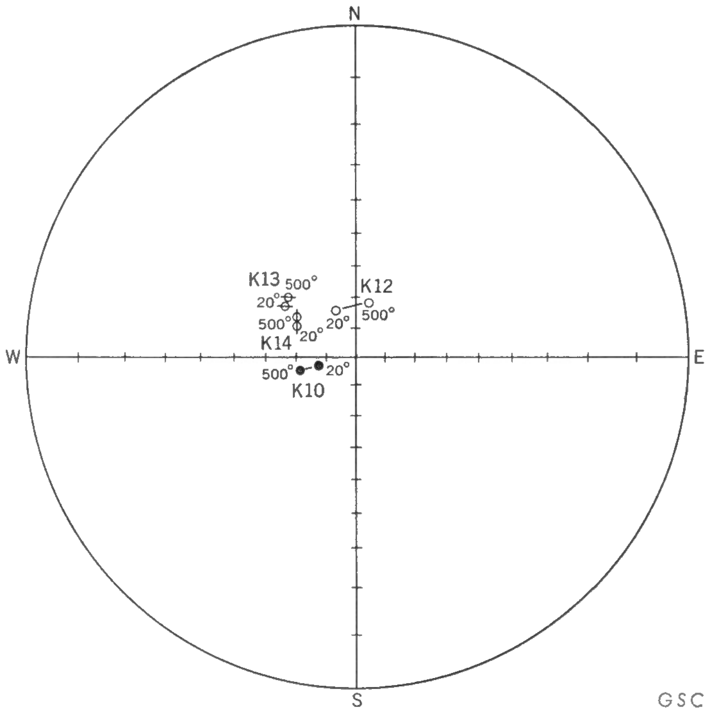


**Figure 3b.** *Diagram showing absence of significant change in direction of remanent magnetization with rise in temperature, degrees Centigrade, in Chequamegon sandstone (not corrected to bedding plane).*

remanence. As the intensity of remanence decreased, there was, however, no change in direction of the magnetic vector.

Figure 3 shows the thermal decay of some samples from the Chequamegon sandstone. These experiments were undertaken to test whether the Chequamegon sandstone was stably magnetized. As its natural remanence shows the Curie type decay (*see* Fig. 3A), and there was no change in the direction of the magnetic vector (*see* Fig. 3B), the author believes that the magnetization of the Chequamegon is stable.

## Alternating Field Demagnetization

Another method that can sometimes be used to eliminate an unstable isothermal component is to subject the rock sample to an alternating magnetic field and then slowly to reduce this field to zero. Two precautions are necessary. Firstly, there must be no steady magnetic field superimposed on the alternating one, because some rocks can acquire what Thellier (1946) called an anhysteretic magnetization, which is strong and stable. To prevent the introduction of such an anhysteretic component during an A. C. demagnetization it is necessary to cancel the earth's magnetic field by means of Helmholtz coils. Certain rocks, however, can acquire very strong anhysteretic magnetization even in very small steady fields, and in such cases they will pick up an appreciable anhysteretic component if the earth's field has not been perfectly cancelled. Therefore, with some rocks it is impossible to cancel the earth's field sufficiently well, and such rocks cannot be demagnetized successfully by the A. C. method. A second precaution is that the A. C. field must be reduced very smoothly. Any sudden changes in the alternating current cause transients, which can introduce spurious magnetizations. Experience has shown that most current regulators, such as variable resistors and variable transformers, are too discontinuous for this purpose. This problem is solved by slowly withdrawing the A. C. coil from the sample.

A number of samples reported in this paper were subjected to A. C. demagnetization in order to eliminate their unstable components. In some cases these attempts were successful, but in others random and spurious magnetizations were introduced, probably as a result of imperfect cancellation of the earth's magnetic field and the consequent anhysteretic magnetization.

Figures 4 and 5 show the effect of A. C. demagnetization on samples of rocks from the Duluth gabbro and Logan diabase dykes respectively. These rocks were subjected to A. C. fields of about 200 oersteds, and it can be easily seen how the magnetization moved from the original direction towards the mean direction of each group. A. C. demagnetization was tried on other rocks from the Lake Superior region, but was unsuccessful apparently because the process introduced new components of magnetization.

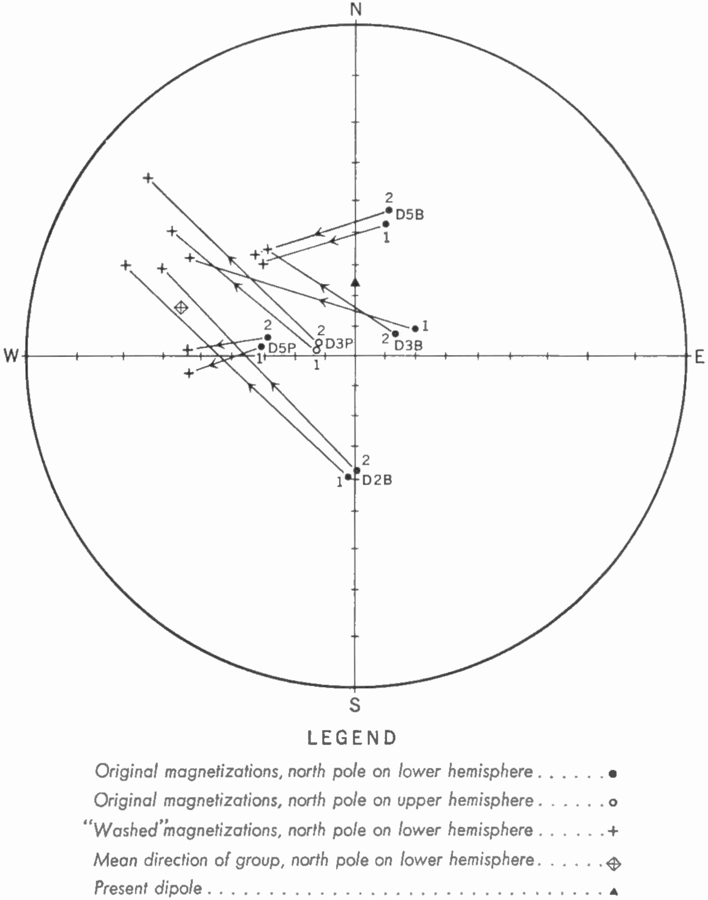


**Figure 4.** A.C. demagnetization of Duluth gabbro.

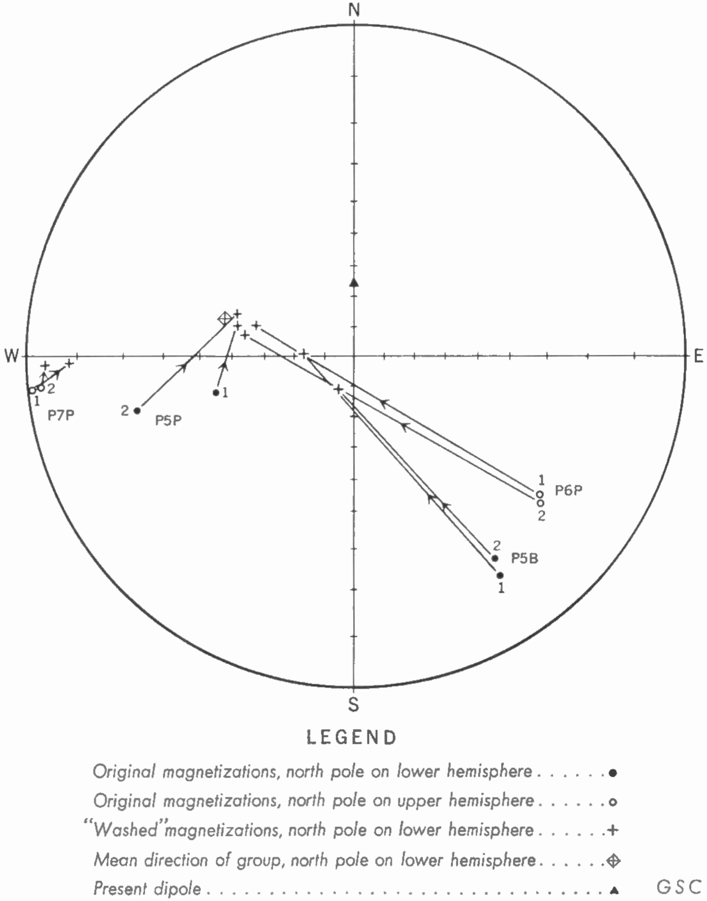


**Figure 5.** A.C. demagnetization of Logan diabase dykes.

# *Chapter III*

# GEOLOGY

Before the palaeomagnetic results for the Lake Superior region are presented and their interpretation discussed, it is useful to review briefly the geological setting of the area.

Because of the importance of the copper deposits in the trap rocks of the Keweenaw Peninsula, a great deal of geological work has been done by United States state and federal surveys, and Canadian geological surveys on the Keweenawan Series. The following general description is drawn largely from Monograph 52 of the United States Geological Survey (Van Hise and Leith, 1911), from *Sandstones of the Wisconsin Coast of Lake Superior* by F. T. Thwaites (1912), from *Bedrock Geology of the Ahmeek Quadrangle, Michigan* by White, Cornwall, and Swanson (1953), and from *Fort William and Port Arthur, and Thunder Cape Map-areas, Thunder Bay District, Ontario* by Tanton (1931). Additional information about the Keweenawan in the United States was obtained from papers by Butler and Burbank (1929), by C. K. Leith, R. J. Lund, and A. Leith (1935), by Sandberg (1938), and by G. M. Schwartz (1949). The Canadian Keweenawan is well described in papers by E. S. Moore (1927), by R. G. McConnell (1927), by E. W. Nuffield (1956), and by J. E. Thomson (1954). The nature and petrology of the rocks, particularly of the traps, have been well described by Broderick (1935), and by Cornwall (1951a, b).

## Keweenaw Peninsula

The Keweenawan forms a very interesting series of deposits in the Lake Superior region, which are remarkable for their great thickness and extensive volcanism. Structurally, the Keweenawan rocks are in a large syncline that forms most of the Lake Superior basin (*see* Fig. 6). Therefore, in most localities along the shore of Lake Superior, the Keweenawan, wherever it is found, dips towards the centre of the lake. The dips range from the vertical to a few degrees from the horizontal. Mostly the lower beds dip more steeply than the overlying beds. In its type locality on the peninsula of the same name the Keweenawan consists of about 11,000 feet of exposed lava, overlain by 4,500 feet of conglomerate that contains a certain amount of sandstone and lava. This conglomerate is in turn overlain by shale and sandstone, of which about 3,000 feet are exposed on the peninsula. The lavas are now known generally as the Portage Lake, the conglomerate as the Copper Harbor, the shale as the Nonesuch, and the sandstones as the Freda. The Portage Lake lavas have been assigned to the middle Keweenawan, whereas the other formations mentioned above are considered to be upper Keweenawan. The lavas dip towards the lake at angles of from 30 to 60 degrees, the conglomerate 20 to 30 degrees, and the Freda and Nonesuch 0 to 20 degrees.

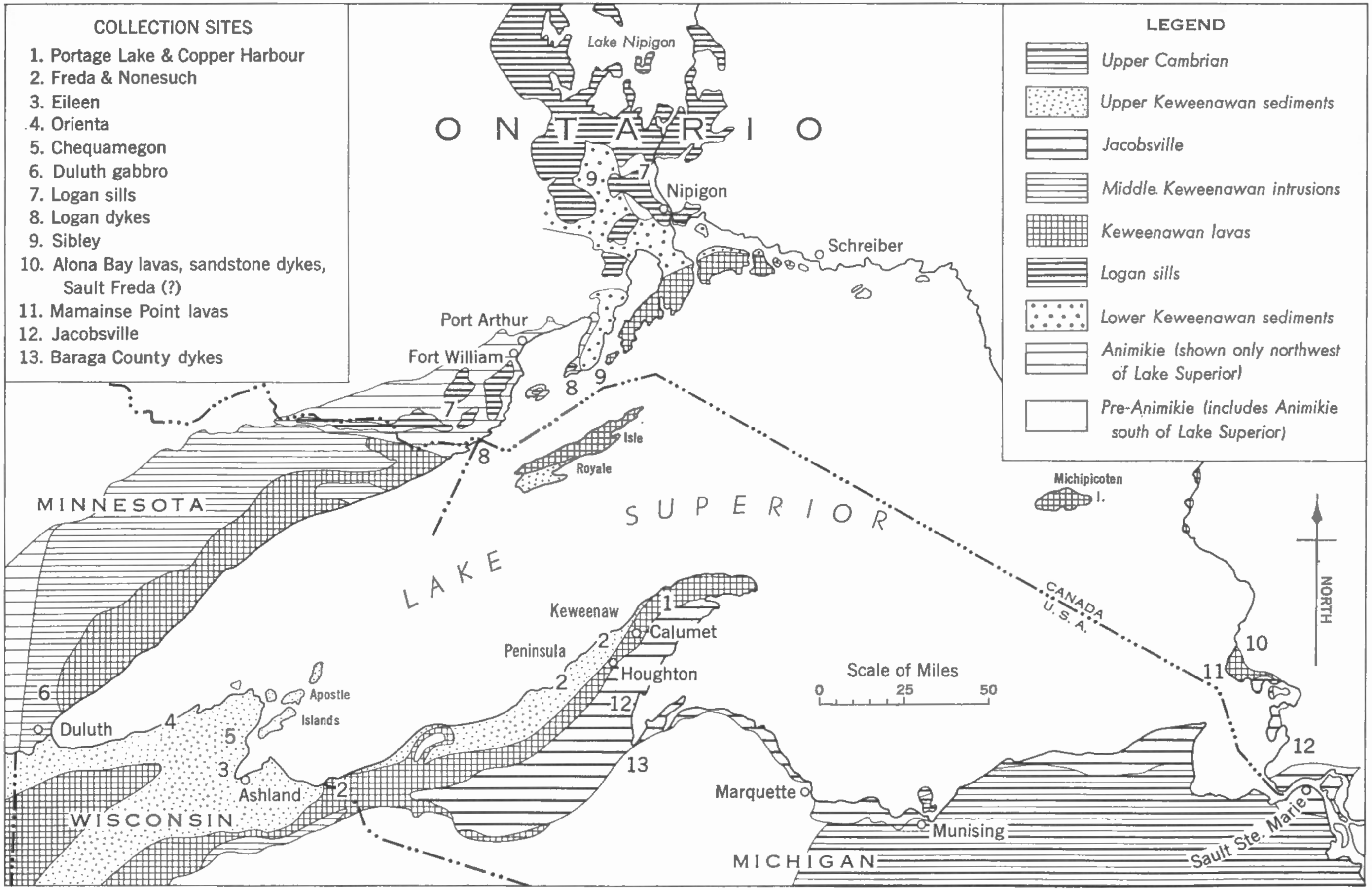


*Figure 6.* Geology of Lake Superior region as interpreted by Van Hise and Leith (1911), and Hamblin (1958).

The Keweenaw Peninsula protrudes into Lake Superior from the Upper Peninsula of Michigan, near its base it trends to the northeast but towards the tip gradually swings around to the east and slightly south of east. This shape reflects the change in the strike of the Keweenawan traps from northeast to east-southeast. Topographically, the peninsula on its northwestern and northern sides has a long continuous hogback ridge due to the outcropping of the Keweenawan trap. There are, in places, occasional notches in this ridge where faults cross it and create zones of weaker rock that are more easily eroded to form small valleys. East and south of this ridge is a flat lowland that is underlain by at least 2,000 feet of flat-lying Jacobsville sandstone.

The Portage Lake lavas range in composition from basalts to andesites with a few rhyolites. The lavas are generally of the order of a few tens of feet thick although the 'Greenstone' flow is known to be as much as 1,500 feet thick. Interbedded with the lavas are occasional thin layers of conglomerate. These conglomerate beds are very important, because they form persistent and easily distinguishable stratigraphical horizons and because many are the sites of important copper deposits.

The Copper Harbor conglomerate consists of boulders and pebbles from a few inches to a foot in diameter embedded in a coarse, red, sandy matrix. About 90 per cent of the boulders are rhyolites or syenites whereas the remaining 10 per cent are basalts and andesites of the same type as the underlying Portage Lake lavas. There are, within the Copper Harbor, lenses of coarse reddish sandstone, which is poorly sorted and generally crossbedded. The reason that more acidic rocks prevail among the boulders and pebbles of the Copper Harbor conglomerate is probably because they were much more viscous when extruded than the basalts, and andesites formed hummock-like masses that were easily eroded and broken up to form boulders for the conglomerate. Near Calumet, Michigan, two lava flows are exposed towards the centre of the Copper Harbor and at the top about ten flows occur with an aggregate thickness of 500 feet (White, Cornwall, and Swanson, 1953). The Copper Harbor apparently lies conformably on the Portage Lake.

There is a sharp break in rock type at the top of the Copper Harbor. Within a few feet, the conglomerate with boulders and cobbles gives way to the fine-grained siltstones of the Nonesuch shale unit. Thus, although the two appear to be conformable, it is possible that a long period of time elapsed between the end of the deposition of the Copper Harbor sediments and the beginning of the deposition of the Nonesuch. The latter consists principally of reddish, fine-grained sandstone, siltstone, and shale. About 80 miles to the southwest of the Keweenaw Peninsula, the Nonesuch is represented by a black shale in which important copper mineralization has taken place. The Nonesuch is a relatively thin formation, being only 700 feet thick.

The Nonesuch grades upward into the Freda through a banded fine-grained sandstone that is known locally as 'tiger stone'. The bands in the 'tiger stone' consist of alternate light and dark layers about half an inch thick, which look rather like the stripes of a tiger. The dark layers are due to the higher concentration of iron ore minerals. The Freda itself is also mainly made up of fine- and medium-grained sandstone, siltstone, and shale. The Freda is a typical continental deposit. The colours range from red and purple to grey. Some of the sandstones are crossbedded, and mud-cracks and rain-drop imprints can be found.

On the Keweenaw Peninsula the southeastern boundary of the Portage Lake lavas is marked by the Keweenaw fault, which is a high-angle thrust fault along which the lavas have been thrust to the east. It is thought that the main activity of this fault was during Keweenawan time, but as the Jacobsville sandstone, which is on the eastern (footwall) side of the fault, has been crumpled along their mutual contact there must have been some movement in post-Jacobsville time. Some smaller faults, principally tear faults, were observed cutting the trap obliquely to the strike of the lava flows, but they are relatively unimportant.

Southeast of the Keweenaw fault the flat-lying Jacobsville sandstone consists principally of fine-grained, red, buff, and white sandstones. The red sandstone contains white spots, where evidently the hematite, which gives the rock its distinctive colour, has been reduced perhaps by the presence of organic material. The base of the Jacobsville is not visible on the peninsula, but farther to the southeast near L'Anse it is observed to rest on folded Animikie-type rocks.

## Northern Wisconsin

Formations of the Keweenaw Peninsula can in general be projected to the southwest along strike towards the Michigan-Wisconsin boundary. The lavas and overlying sediments dip towards the lake and are cut off to the east by a large thrust fault. To the east of this fault are flat-lying Jacobsville sandstones.

Near Gogebic Lake a branch of the Keweenawan lavas juts out into the upper Michigan Peninsula on an east-west line. Along the Montreal River, which forms the Wisconsin-Michigan boundary, the strata are almost vertical, but to the northwest, near Ashland, Wisconsin, the uppermost Keweenawan is flat lying. Thwaites (1912) studied the section near Ashland and found that the Freda sandstone, which is the uppermost Keweenawan unit on the Keweenaw Peninsula, at Ashland is overlain by 11,000 feet of sediments. The geological column, deduced by Thwaites from his studies of sections along the Montreal River and near Ashland, is given in Table I.

Table I

| Bayfield Group | Thickness (feet) |
|---|---|
| Chequamegon sandstone: red and white sandstone composed predominantly of quartz grains | 1,000 |
| Devil's Island sandstone: pink and white, pure quartz sandstone | 300 |
| Orienta sandstone: like the Chequamegon, but with more feldspar | 3,000 |
| Oronto Group | |
| Amnicon formation: red and greenish shales, arkose, sandstone, and conglomerate | 5,000 |
| Eileen sandstone: red and white arkose and sandstone | 2,000 |
| Freda sandstone: fine-grained, red and greenish, arkosic sandstones | 12,000 |
| Nonesuch shale: black, grey and red arkose and shale | 350 |

In this region the Keweenawan has been folded as well as tilted, and to the northwest of Ashland the Douglas fault has thrust the traps northwestwards onto the highest sediments, including the Chequamegon sandstone. This fault, like the Keweenaw fault, is probably a surface manifestation of the general structural forces that created the Lake Superior syncline, and thus most of the faulting probably took place during or shortly after Keweenawan time. As Upper Cambrian structure is not affected by these faults and reflects little of the Keweenawan structure, the main faulting can be presumed to be pre-Upper Cambrian, and consequently the whole of the Keweenawan is pre-Upper Cambrian (Atwater and Clement, 1935). East of the Keweenaw Peninsula rocks of the Jacobsville facies have been found lying with a small angular unconformity under the Upper Cambrian Munising sandstone at Grand Island, Michigan (Hamblin, 1958).

From the existence of strongly positive gravity anomalies and the evidence of a few bore-hole cores, it appears that the Keweenawan probably extends far to the southwest of Lake Superior, perhaps as far as the Nebraska-Kansas border, under a cover of younger sediments (Thiel, 1956).

## Northeastern Minnesota

In northeastern Minnesota all three units, lower, middle, and upper, of the Keweenawan are supposed to be represented. The Keweenawan rocks outcrop in two areas, whose structural relationships with one another are not well known although they are adjacent. The most important region in which lower and middle Keweenawan are represented, is a great crescent-shaped area, with its concave side towards Lake Superior, extending from Duluth to the Canadian boundary.

The lower Keweenawan in northeastern Minnesota is represented by the Puckwunge conglomerate, which is visible in various places near Grand Portage Bay near the Canadian border where it lies at the top of the Animikie group.

The conglomerate grades up into sandstone and is overlain by the middle Keweenawan lavas. On St. Louis River near Duluth, a quartzite conglomerate about 100 feet thick is exposed which is conformable below the Keweenawan lavas and contains some pebbles of Keweenawan rocks. Winchell (1899) regarded this conglomerate as equivalent to the Puckwunge, but more than 100 miles of terrain (in which little outcrop is exposed and the Duluth gabbro appears) separates the two localities, and no definite correlation can be made. Furthermore, although the Puckwunge conglomerate is considered to be lower Keweenawan, it is separated from the lower Keweenawan sediments of the Sibley Peninsula and Lake Nipigon area farther to the northeast in Ontario, and on geological field evidence it is impossible to say whether the sediments in Minnesota and in Ontario are of the same age.

In northeastern Minnesota the most important Keweenawan rocks are those of the middle unit, which can be divided into two great divisions (1) the extrusive lavas with associated interbedded sediments, and (2) the intrusive rocks, of which the greatest is the Duluth gabbro. The lavas are mostly a series of well-stratified basic flows with gentle south-eastward or lakeward dips, similar in petrology to the Portage Lake lavas. Associated with the basic lavas are minor amounts of more acidic lavas that are quartz porphyrites and felsites.

Interbedded with the lava flows are thin beds of conglomerate and sandstone, which rarely exceed a few feet in thickness. Sandberg (1938) found that, in his classic section along the shore of Lake Superior near Duluth, interbedded fragmental rocks total a mere 267 feet out of a total thickness of 20,539 feet for the Keweenawan lavas.

Intruded into the Keweenawan lavas are a great many sills of diabase, some of which are difficult to distinguish from the lava flows themselves. The Lester River and Endion sills reach a thickness of 1,000 feet each, and there are many thinner sills. The total thickness of these diabase sills is about 4,000 feet.

The largest intrusion in the region is the Duluth gabbro, which has been classified as a lopolith by Grout (1918). Within this tremendous intrusive body are found many different rock types, ranging from a very magnetitic gabbro through anorthosite to a red granitic rock. Texturally the gabbro varies from a very coarse rock to one that is practically aphanitic. In most localities the gabbro is massive, but in many places, especially near its base, it has a layered structure. In the Duluth area the gabbro is 14,500 feet thick and is intruded between the Keweenawan lavas. Most of the lavas lie above the gabbro, but some 2,500 feet of lava intervenes between the base of the gabbro and the top of the Carlton slate. Farther to the north and northeast of Duluth the base of the gabbro cuts successively down through the Keweenawan lavas and Animikie sediments to lie directly on the crystalline basement near Babbitt, Minnesota. Still farther to the northeast it cuts back up through the Animikie and pinches out as it approaches Lake Superior. The gabbro may reach a maximum thickness of more than 50,000 feet between Lake Superior and the Vermillion Iron Range.

To the south and southeast of Duluth in Douglas county, Wisconsin and Pine county, Minnesota, is a band of basic lava flows that strikes northeast and dips southeast. This band of lavas has been thrust northwestwards over the upper Keweenawan sandstones. The latter is an extension to the southwest of the sandstones of the Bayfield group, found near Ashland, Wisconsin, and described in a previous section. In the metropolitan Duluth area the red Fond du Lac sandstones and shales are present in the St. Louis River valley. Thwaites (1912) correlated these rocks with the Oronto group (*see* Table I). Later, Atwater and Clement (1935) correlated the Fond du Lac with the Amnicon, which is the upper part of the Oronto, and still later Tyler, Marsden, Grout, and Thiel (1940), on the basis of their accessory mineral studies, decided that the Fond du Lac is equivalent to the lower part of the Orienta. Whatever the exact correlation for the Fond du Lac beds is, it seems certain that these beds are of upper Keweenawan age. A more difficult problem with the Fond du Lac concerns its relations with the Duluth gabbro and Keweenawan lavas, which outcrop with very great thickness just 2 miles north of St. Louis River. There are three possible explanations (1) the flows and gabbro thin drastically in the 2-mile interval; (2) a fault intervenes between the two rock types; (3) an unconformity of considerable proportions exists between the lavas and gabbro, and the Fond du Lac. Explanation (1) is very improbable because of the great thickness of the igneous rocks only 2 miles from the Fond du Lac beds. Explanation (2) is plausible, but there is little direct evidence for such faulting. Similarly, there is so far little direct evidence for explanation (3).

## Northwestern Ontario

On the northern shore of Lake Superior are the lower members of the Keweenawan series. The Keweenawan in this region and the Animikie have been grouped together by Tanton (1931) to form his Kaministikwan group. These rocks occupy a zone about 25 miles wide along the north shore of Lake Superior for 180 miles, from Gunflint Lake to the vicinity of Schreiber, and extend north over a large area including Lake Nipigon.

Tanton divided the Animikie into three formations, which are in ascending order, the Kakabeka, the Gunflint, and the Rove. As this report is not concerned with the Animikie directly, it is sufficient to say that the Animikie consists of about 2,000 feet of conglomerate, sandstone, ferruginous chert rocks or iron-formation, greywackes, and thinly bedded black shales. These rocks dip at low angles to the southeast under Lake Superior.

Overlying the Animikie are the sediments of the Sibley series of lower Keweenawan age, which consist principally of red, fine-grained sandstones and siltstones and, although the two formations generally appear to be conformable at individual localities, regional studies show that in some places the Sibley overlaps the Animikie and rests on the early Precambrian basement and that the two formations are in fact disconformable.

In some places the basal conglomerate of the Sibley has many angular fragments of Animikie iron-formation. This fact lends support to the interpretation that a period of erosion separated the Sibley from the Animikie. The type locality of the Sibley series is on the Sibley Peninsula near Port Arthur, but farther to the northeast large areas around Lake Nipigon are underlain by almost flat-lying red sediments. These rocks have been described by Wilson (1910) in ascending order as a basal conglomerate, red sandstone and shale, and red dolomites; they have a total thickness of about 550 feet. On geological evidence it is probable that these sediments are equivalent to the Sibley series.

The Sibley is overlain by the Osler series, which is a succession of lavas and sediments. The lower contact of the Osler on the Sibley is marked by a basal conglomerate, which contains pebbles derived from the underlying Sibley. This conglomerate is succeeded by beds of sandstone and red mudstone with a total thickness in some places of 200 feet, and these sediments are in turn overlain by a thick succession of lavas that contain a few interbedded sediments. Tanton has found structures at the contact of the lowest lava flow with the underlying sediments that he interprets as indicating that the sediments were plastic when the lava flowed over them and that, therefore, no considerable time elapsed between the deposition of the sediments and the overlying lava. The lavas are lithologically uniform basalts similar to the middle Keweenawan lavas of Michigan, across the lake, and there is evidence that they flowed from the centre of Lake Superior northwestwards towards the margin.

The Kaministikwan sediments are intruded by dykes and sills of diabase and related rocks. In two main areas the diabase sills are extensively developed. Around Lake Nipigon a very large area is underlain by diabase sills, which intrude the red clastic sediments and dolomites that are supposed to be equivalent to the Sibley series. Around Thunder Bay and southwest of Port Arthur to Pigeon River is another area where diabase sills intrude the Animikie sediments extensively. The sill capping the Sleeping Giant at Thunder Cape is known to intrude the lower part of the Sibley series as well as the Animikie, but in general the Sibley in its type locality is mainly cut by diabase dykes. Near the mouth of Pigeon River occur numerous prominent diabase dykes with northeasterly strikes. Similar dykes, which cut the Animikie and Sibley, are present on the southern end of Sibley Peninsula, and are probably the northeastward continuation of the dykes near Pigeon River. Farther to the northeast Tanton reported that dykes of diabase also cut the Osler strata, and on St. Ignace Island and the nearby mainland the lavas are intruded by diabase sills and quartz porphyry.

Many geologists have assumed that these extensive intrusions were all emplaced during the same geological period and probably came from a parent body of enormous dimensions. The close resemblance between the diabase intrusions and the lavas of the Osler series implies that they both came from the same magma source and were formed at roughly the same time. However, Tanton showed that the intrusive rocks were not all injected at the same time, because he found places where one intrusion with chilled edges cuts another, indicating that sufficient time had passed to allow the earlier intrusion to cool. Furthermore, Grout (1926), from his studies of the contact metamorphism of the Duluth gabbro concluded, "The diabase sills in the Animikian rocks are definitely older than the (Duluth) gabbro, and some may be older than the Keweenawan". As it is known that some lava was extruded during Animikie time, it is possible that some of the dykes and sills that do not intrude the lower Keweenawan sediments are the intrusive phase of this vulcanism.

## Eastern Lake Superior

Although much of the east half of the Lake Superior basin is probably underlain by Keweenawan rocks, exposures are much less numerous and the geological knowledge consequently much less perfect. The principal areas of Keweenawan rocks are on Michipicoten Island and along the eastern shore of Lake Superior in Ontario north of Sault Ste. Marie.

No lower Keweenawan is definitely known to be exposed in eastern Lake Superior. Moore (1927) regarded a vast number of diabase dykes near Montreal

River north of Sault Ste. Marie as lower Keweenawan, because they do not seem to cut lavas that are supposed to be of middle Keweenawan age. These dykes have been sheared in places, and in certain shear zones where they are in contact with pegmatite and pegmatitic gneiss the diabase has been mineralized with uranium. On purely geological evidence it is impossible to say how much time intervened between the intrusion of the dykes and the extrusion of the overlying lavas.

Middle Keweenawan lavas are present on Michipicoten Island and at various isolated localities on the eastern shore of Lake Superior, including Cape Choyne, Cape Gargantua, Alona Bay, Mamainse Peninsula, and Gros Cap. On Michipicoten Island the lavas strike east-west and dip at from 55° to 15°S. The lower lavas dip more steeply than the upper, and this lessening of the dips is very similar to the situation on the Keweenaw Peninsula. On the eastern shore the lavas strike north-south and dip at 25° to 60°W, into the lake. Thus, if the lavas on Michipicoten Island are continuous with those on the eastern shore, there must be a sharp inflection in the strike of these rocks near Michipicoten Harbor.

The most extensive development of the lavas is at Mamainse Point, where Moore (1927) estimated their thickness to be about 16,000 feet. Interbedded with the lavas are beds of conglomerate and sandstone, and they were also intruded by dykes and irregularly shaped bodies of felsite.

Overlying the lavas with considerable unconformity are flat-lying or gently tilted sandstones that apparently occur in two forms. The first type is exposed along the eastern shore of Lake Superior, where small isolated patches of well-bedded, reddish grey sandstone overlie the lavas with considerable unconformity. These sandstones generally have a dip of 15 degrees. Along the south shore of Batchawana Bay the author found an area of highly disturbed sandstone, presumably the result of faulting. In appearance these rocks are much like the Freda sandstones of Michigan. The second type of sandstone is exposed around Sault Ste. Marie, in the valley of St. Mary's River. This sandstone is much more irregularly bedded, poorly sorted, and poorly consolidated than that of the first type. It is also characterized by a much redder colour with white spots and layers, where the hematite appears to have been bleached. Its physical appearance closely resembles the Jacobsville sandstone. This sandstone also appears to be much more extensive and thicker than the sandstones of the first type. Finally, it appears everywhere to be nearly flat lying and undisturbed.

McConnell (1927) lumped these sandstones together into one group which he called the Lake Superior sandstone. However, Hamblin (1958) recently suggested that the first type of sandstone should be correlated with the Freda and has shown that the second type, exposed at Sault Ste. Marie and around Whitefish Bay, is an eastward extension of the Jacobsville sandstone of Michigan. The latter is definitely younger than the Portage Lake lavas, on which it rests near the South Range, Michigan, with a considerable angular unconformity. Hamblin (op. cit.) suggested that certain pebbles found in the Jacobsville are from the Freda, to which they bear a strong lithological similarity, implying that the

Jacobsville is younger than the Freda. The Jacobsville is overlain unconformably by the Munising formation of Upper Cambrian age. Whether this unconformity represents a major break in sedimentation cannot be determined, but supporting the idea of a considerable time break are the facts that certain clastic dykes, which fill tensional cracks at the top of the Jacobsville, are truncated by the overlying Munising and that there is a marked change in the heavy mineral suite between the Jacobsville and Munising.

In the lavas along the shores of Alona Bay are clastic dykes that appear to be the result of tension cracks being filled by medium-grained, red sandstone. These dykes cut not only the lavas themselves but also the veins of calcite that accompanied the copper mineralization.

## General Geological History of the Keweenawan

The sequence of events for Keweenawan time can be summed up as follows. A syncline, perhaps as much as 800 miles long and 100 miles wide, was formed in Animikie or lower Keweenawan time (*see* Fig. 6). It may well have been due to downwarping caused by the compression that also produced the Keweenaw and Douglas thrust faults. Some sediments were laid down in this basin, but were followed by enormous outpourings of lava. Basic lavas welled up quietly from fissures in the centre of the syncline, filled it, and overflowed to the borders of the syncline. The more acidic lavas being very viscous formed hummock-like hills, which were eroded to form boulders in the conglomerates interbedded with, and overlying the lavas. The extrusive phase ceased and downwarping continued. Already active thrust faults raised the lavas along the margins of the syncline, thus exposing them to erosion. The detritus from the lavas and the adjacent pre-Keweenawan rocks was carried by streams and deposited in the syncline. Studies of current crossbedding show that the streams flowed from the margins towards the centre of the syncline. Sedimentation kept pace with the downwarping and the surface of the syncline was always shallow as indicated by the ripple-marks, mud-cracks, rain-drop imprints, and the well-oxidized condition of the sediments. During the process of sedimentation and volcanic extrusion, basic intrusions such as the Duluth gabbro were emplaced. Then ensued a long period of erosion which created a peneplain, and during which the Keweenawan syncline ceased to be active. This was followed by Upper Cambrian sedimentation.

## *Chapter IV*

# DESCRIPTION OF RESULTS

## Portage Lake Lavas

The Portage Lake lavas are a thick sequence of basaltic and andesitic flows with a few thin interstratified layers of rhyolitic conglomerates. They vary both in composition and in grain size. Some of the lavas are uniformly fine grained, but most increase in grain size towards the centre. The central massive lava is generally capped by a layer of amygdaloidal lava, which may be fragmental and have many vesicles.

Most of the massive lava is exceedingly fresh and shows few signs of alteration and metamorphism after solidification. However, it appears that the iron ore minerals may have undergone considerable alteration before the lava finally solidified. Butler, *et al.* (1929) showed that the principal alterations, aside from copper ore deposition and surface weathering, occurred in the magnetite and olivine, the pyroxene and feldspar generally being unaffected. The olivine, which at the time of crystallization was common in many of the lavas, is rarely preserved and has been replaced by serpentine and hematite. Similarly, the magnetite has been altered more or less completely to hematite. In cases where hematite has replaced magnetite completely, irregular stringers of hematite extend from the magnetite grain among the surrounding minerals. This suggests that in the alteration of magnetite a volume increase forced some of the hematite to lodge outside the original boundary of the magnetite. Some of the magnetite was evidently titaniferous, for its alteration has produced numerous granular particles of a non-metallic mineral, probably titanite, as well as the more abundant hematite. It is extremely important to note (Butler and Burbank, 1929) that the alteration of these two minerals, so far as can be told, was almost magmatic. In other words, it took place very shortly after the rock had solidified unless these early iron-bearing minerals were already altered by the time the pyroxene and the feldspar had crystallized around them. In either case the iron ore minerals probably acquired their magnetization at solidification or just after.

The hematite itself is partly in small ragged lumps but occurs mostly in very small, short crystalline plates distributed through the opaque matrix. The hematite in places is so abundant and fine grained as to give a reddish colour to the lavas, especially in the amygdaloidal tops. All in all, the particles of hematite form a pattern that seems characteristically igneous. They are not deposited in vesicles nor included in feldspar microlites but are crowded around both as if pushed aside by, or attracted to, the growing crystals and expanding bubbles. This evidence suggests that the hematite was formed before the solidification of the rock and is not a product of late alteration by fumaroles.

A total of thirty-one samples of the Portage Lake lavas (*see* Fig. 6, site 1) was collected for palaeomagnetic measurements. Fairly complete descriptions of the localities where these samples were collected are given in Table II. Because of the difficulty in measuring the strike and dip of the lavas in isolated outcrops, they have been computed from the geological maps and cross-sections compiled by the United States Geological Survey. These maps were also used to locate the stratigraphic height of each locality above the St. Louis conglomerate. A total of about 1,850 feet of lava were sampled in this collection.

The results for each individual sample corrected for geological dip are given in Table II and are plotted in Figure 7; the mean results are shown in Table XVIII.

## Table II

*Results of Portage Lake Lavas*

| Sample | D | I | Description of locality |
|---|---|---|---|
| KO1 | 281.0 | +30.5 | From a field in the SE ¼ of NE ¼ of sec. 6, T56N, R32W between the main highway and the abandoned right of way of the Copper Range Railroad. Stratigraphic height above St. Louis conglomerate 5,400′; approximate thickness sampled 300′; strike N34°E, dip 41°W. |
| KO4 | 287.5 | +38.0 | |
| KO5 | 280.0 | +37.5 | |
| KO6 | 282.5 | +46.0 | |
| KO8 | 284.5 | +29.0 | |
| KO9 | 270.5 | +55.5 | |
| KO10 | 289.0 | +36.0 | |
| KO11 | 302.0 | +29.5 | |
| KS1 | 286.5 | +45.0 | From lavas exposed on hill just to the northwest of the Allouez Conglomerate Mine in the NE ¼ of the SW ¼ of sec. 31, T57N, R32W. Stratigraphic height above St. Louis conglomerate 8,100′; approximate thickness sampled 100′; strike N35°E, dip 39°W. |
| KS2 | 281.0 | +28.5 | |
| KS4 | 281.0 | +36.0 | |
| KT1 | 291.0 | +46.5 | From a site beside Scales Creek near the border between the NW ¼ and SW ¼ of sec. 8, T56N, R32W. Stratigraphic height above St. Louis conglomerate 1,950′; thickness sampled 350′; strike N30°E, dip 43°W. |
| KT2 | 288.5 | +45.0 | |
| KT3 | 284.5 | +57.5 | |
| KT4 | 285.0 | +54.0 | |
| KT5 | 286.0 | +55.0 | |
| KT6 | 289.5 | +54.5 | |
| KT7 | 279.0 | +33.0 | |
| KT8 | 290.0 | +43.0 | |
| L1 | 283.0 | +39.5 | In the city of Houghton; L1 from exposure on corner of Portage and South Streets, L2 and L3 from corner of Isle Royal and South Streets. Stratigraphic height above St. Louis conglomerate 3,100′; thickness sampled 300′; strike N28°E, dip 57°W. |
| L2 | 289.0 | +54.5 | |
| L3 | 276.5 | +46.5 | |
| L4 | 294.5 | +37.5 | In the city of Houghton; L4 and L5 from two flows at corner of Quincy and South Streets, L6 and L7 from two flows at corner of First and South Streets, L8 and L9 from corner of Second and South Streets. Stratigraphic height above St. Louis conglomerate 4,300′; thickness sampled 700′; strike N27°E, dip 57°W. |
| L5 | 100.5 | −30.5 | |
| L6 | 302.5 | +38.0 | |
| L7 | 101.5 | −37.5 | |
| L8 | 278.0 | +38.5 | |
| L9 | 269.5 | +36.0 | |
| L10 | 298.0 | +49.0 | In the city of Houghton at corner of Memorial Road and Dakota Street. Stratigraphic height above St. Louis conglomerate 7,500′; thickness sampled 100′; strike N20°E, dip 51°W. |
| L11 | 279.0 | +43.5 | |
| L12 | 261.5 | +35.0 | |

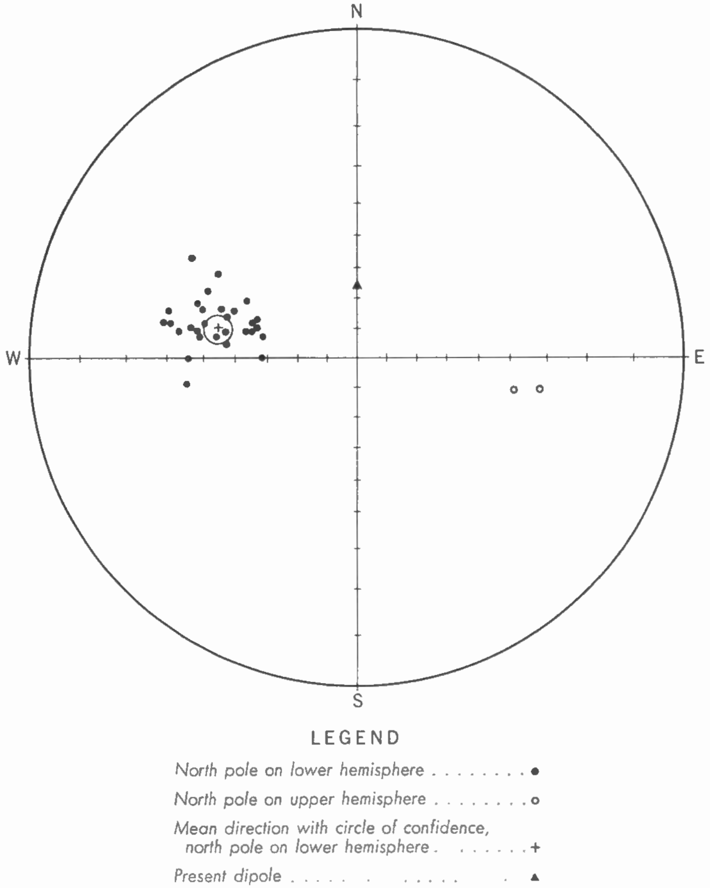


Figure 7. **Directions of magnetization for the Portage Lake lavas.**

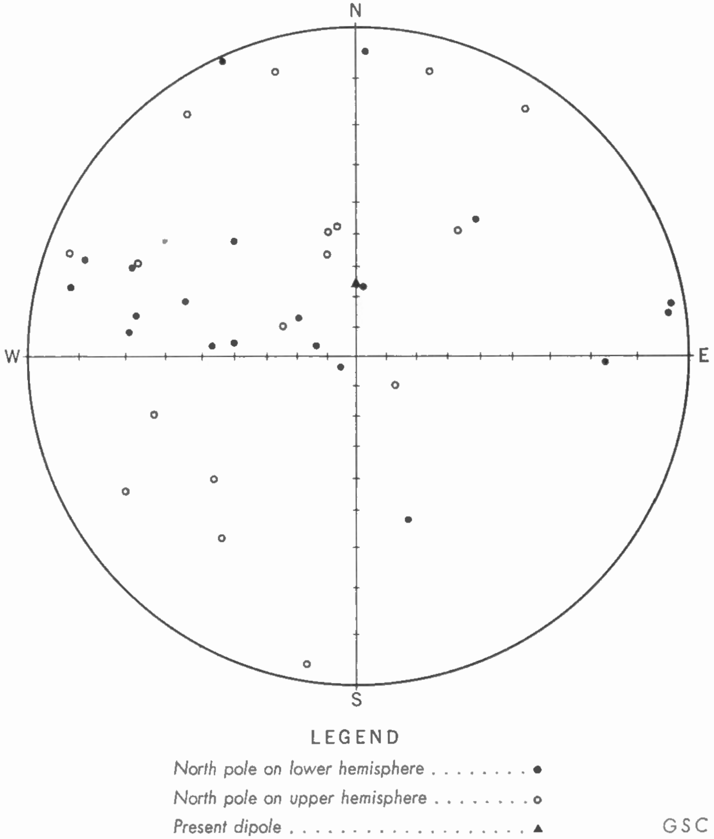


Figure 8. **Directions of magnetization of lava pebbles from the Copper Harbor conglomerate.**

All the results for the Portage Lake lavas are relatively consistent with one another except for specimens L5 and L7. These two rocks have magnetizations that are almost exactly opposite to the mean direction for the Portage Lake lava as a whole. L5 and L7 are therefore reversely magnetized, but whether this reversal is due to a reversal of the earth's magnetic field at the time of their formation, or whether it has been produced by one of the Néel mechanisms, it is impossible to say.

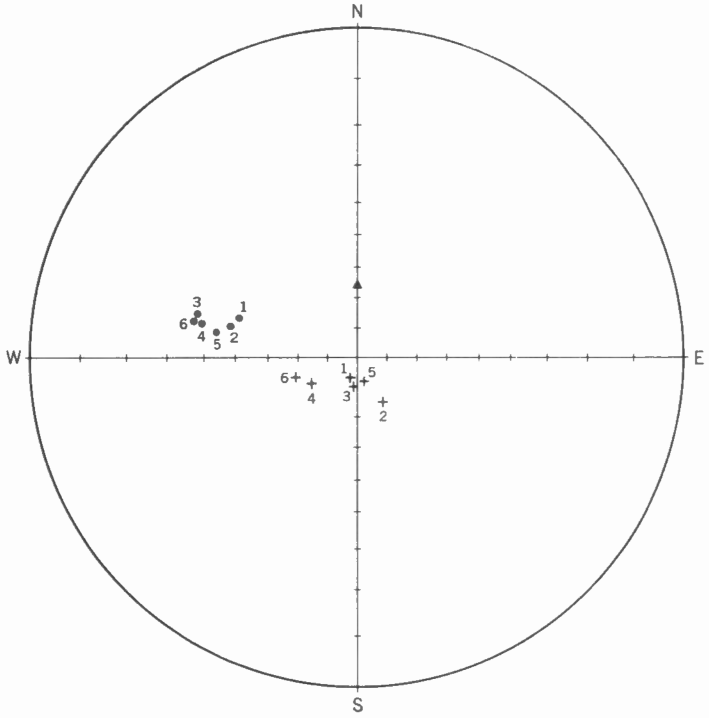
In order to test the magnetic stability of the Portage Lake lavas thirty-six cobbles of basalt, probably derived from the Portage Lake lavas, were collected from the overlying Copper Harbor conglomerate and measured. If these lava cobbles had acquired a stable magnetization when they were cooled through the Curie point and before they were eroded to form cobbles, the direction of the magnetization in the cobbles in the conglomerate should be randomly oriented. A stereographic plot of these results shows this to be so (Fig. 8). This dispersion would thus indicate considerable magnetic stability. It is worth noting that the residual mean direction of magnetizations for the cobbles ($D=309.0$, $I=+16.5$) is fairly close to that of the Keweenawan as a whole and quite different from the present earth's field. This may be purely accidental, or it may indicate that the lavas acquired a small post-solidification magnetization during Keweenawan time.

Another method of testing the magnetic stability of the Portage Lake lavas is to find out whether the directions of magnetization corrected for geological dip are more or less scattered than the directions not so corrected. If the rocks are stable, then the corrected directions should have a smaller dispersion than the non-corrected directions. For the purpose of this study the samples of lava have been grouped into six sites, and the directions of magnetization, corrected and not corrected for geological dip have been computed for each site. These results are presented in Table III and Figure 9. It can be seen in the stereogram that the non-corrected directions are more scattered, and this surmise is confirmed by Fisher's statistical analysis, which shows that the K is 234 for the corrected samples and only 49.9 for the uncorrected samples.

### Table III

*Corrected and Uncorrected Directions of Magnetization in Portage Lake Lavas*

| Samples | Corrected measurement | | Dip | | Uncorrected directions | | Reference No. on stereogram |
|---|---|---|---|---|---|---|---|
| | D | I | Strike | Dip | D | I | |
| KT1–KT8 | 286.5 | +48.5 | 30°E | 43°W | 200.0 | +80.5 | 1 |
| L1–L3 | 282.5 | +46.5 | 28°E | 57°W | 153.5 | +72.5 | 2 |
| L4–L9 | 284.5 | +36.5 | 27°E | 57°W | 185.0 | +79.5 | 3 |
| KO1–KO10 | 281.5 | +38.5 | 34°E | 41°W | 239.0 | +70.5 | 4 |
| L10–L12 | 279.0 | +43.0 | 20°E | 51°W | 169.0 | +80.5 | 5 |
| KS1–KS4 | 282.0 | +36.0 | 35°E | 39°W | 251.0 | +67.5 | 6 |



GSC

**Figure 9.** **Directions of magnetization in Portage Lake lavas, corrected and uncorrected for dip.**

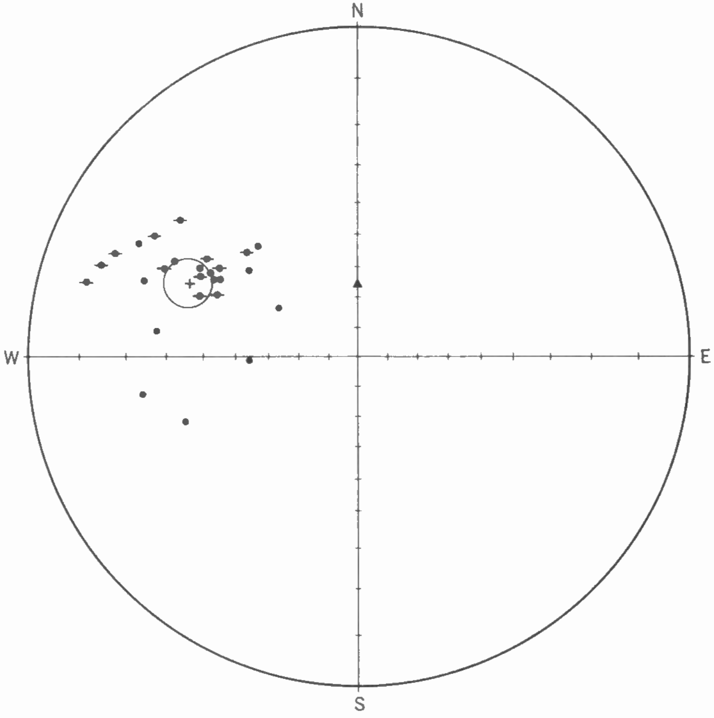


**Figure 10.** **Directions of magnetization for Copper Harbor formation.**

## Copper Harbor Conglomerate

The Copper Harbor conglomerate (White, *et al.,* 1953) apparently conformably overlies the Portage Lake lavas and consists principally of rudely stratified conglomerates. The pebbles and boulders are well rounded to subangular and are chiefly composed of rhyolite. Small quantities of amygdaloidal and massive basaltic fragments can be found, and in some localities they may constitute up to one quarter or one half of the conglomerate boulders and pebbles. On the average, however, the basalt fragments probably constitute only 10 per cent of the total. The matrix of the conglomerate is coarse sand and grit of the same lithological character as the coarser material. Subordinate amounts of medium-grained to gritty sandstone are interbedded with the conglomerate. This sandstone may occur in small lenses a few inches thick or in thicker layers several tens of feet thick. Because it does not outcrop as well as the conglomerate, relative abundance of the two is not known.

Apparently three lava units are interbedded with the conglomerate. The lowermost, which occurs below the middle of the formation, probably consists of two flows, separated by conglomerate. It does not outcrop, but its existence and

**Table IV**

*Results for Copper Harbor Conglomerate*

| Sample | D | I | Description of locality |
|---|---|---|---|
| KH1 | 298.5 | +27.0 | From two lavas exposed near the Allouez-Tamarack Pumping Station road along the boundary between sec. 26 and sec. 35 of T57N, R33W; stratigraphic height above base of Copper Harbor 2,200′; thickness sampled 100′; strike N30°E, dip 31°W. |
| KH2 | 298.5 | +17.5 | |
| KH3 | 308.5 | +45.0 | |
| KH4 | 250.0 | +32.5 | |
| KH5 | 299.0 | +38.5 | |
| KH6 | 299.5 | +33.5 | |
| KJ1A | 295.0 | +25.5 | From sandstone exposed in ditch beside the Allouez-Tamarack Pumping Station road in the middle of sec. 26, T57N, R33W; stratigraphic height above base of Copper Harbor 3,200′; thickness sampled 300′; strike N30°E, dip 30°W. |
| KJ2 | 301.0 | +19.5 | |
| KJ3 | 291.0 | +37.0 | |
| KJ3C | 308.0 | +22.5 | |
| KJ4 | 297.5 | +34.5 | |
| KJ5 | 313.5 | +41.5 | |
| KJ5B | 303.0 | +38.0 | |
| KJ6 | 290.0 | +12.0 | |
| KJ6A | 293.0 | +13.5 | |
| KJ7 | 285.5 | +10.0 | |
| KM1 | 303.5 | +33.5 | From sandstone exposed about ¼ mile west of KH1-KH6, beside road; stratigraphic height above base of Copper Harbor 2,400′; thickness sampled 60′; strike N30°E, dip 30°W. |
| KM3 | 294.5 | +41.0 | |
| KP2 | 302.5 | +59.5 | From lava of 'Outer Trap'; samples collected along section across trap ridge in NE ¼, sec. 34, T57N, R33W; stratigraphic height above base of Copper Harbor 3,800′; thickness sampled 300′; strike N29°E, dip 29°W. |
| KP3 | 277.0 | +28.0 | |
| KP5 | 260.0 | +24.0 | |
| KP7 | 318.5 | +42.0 | |
| KP8 | 299.5 | +39.5 | |
| KP9 | 268.5 | +54.0 | |
| KP10 | 289.5 | +22.0 | |

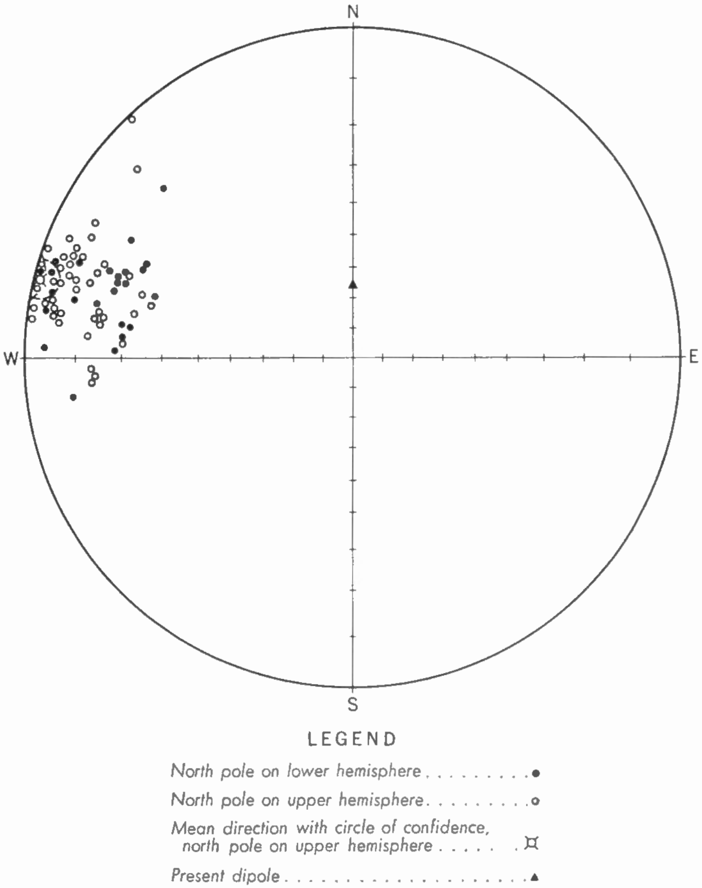


*Figure 11.* ***Directions of magnetization for Freda sandstone and Nonesuch shale.***

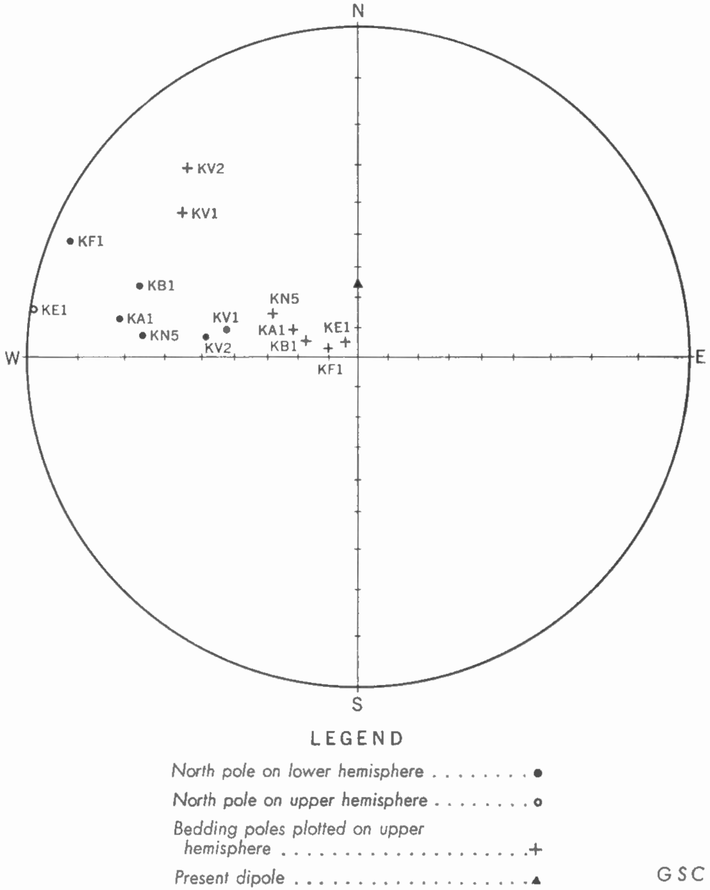


*Figure 12.* ***Directions of magnetization of selected samples of Freda sandstone and Nonesuch shale, not corrected for dip.***

position have been deduced from magnetic surveys, and the lowermost flow has been cut by a drill-hole. The intermediate unit, which occurs at the middle of the Copper Harbor formation, outcrops as two flows separated by conglomerate. The upper unit sometimes called the 'Outer Trap', which occurs at the top of the formation, consists of eight to ten flows that form a hogback ridge near Lake Superior. The individual flows typically are fine-grained andesite, with an amygdaloidal layer 2 to 10 feet thick at the top.

All in all twenty-five specimens, twelve of sandstone and thirteen of lava, were collected from the Copper Harbor formation. Because of poor exposures of the lower half, only the upper half of the formation has been adequately sampled (*see* Table IV and Fig. 10).

## Freda Sandstone and Nonesuch Shale

The Freda sandstone and Nonesuch shale appear to form a conformable unit and are distinguished mainly by minor lithological differences. The upper part of the Nonesuch shale on Keweenaw Peninsula (White, *et al.*, 1953) consists principally of flaggy, ripple-marked, grey to reddish grey siltstone with subordinate amounts of interbedded grey and greenish grey silty shale. The lower part consists chiefly of dark grey siltstone, mostly thinly laminated, with several beds of coarse-grained, very red arkosic sandstone near the base. To the southwest a distinctive black cupriferous shale occurs in the Nonesuch.

On Keweenaw Peninsula the Freda is typically red, fine- to medium-grained sandstone with minor amounts of red shale and conglomerate. The lower part consists of light grey, fine- to medium-grained sandstone with purple streaks. At the contact with the Nonesuch, is the purple banded rock known as 'tiger stone'. Farther to the southwest, along Montreal River, the Freda is coarser, more arkosic, and redder.

All in all sixty-eight samples of rock were collected from the Freda sandstone and Nonesuch shale and measured for remanent magnetism (*see* Table V and Fig. 11).

### Table V

*Results for Freda Sandstone and Nonesuch Shale*

| Sample | D | I | Description of locality |
|---|---|---|---|
| KA1 | 279.5 | − 5.0 | From stream bed of Brewery Creek near Tamarack Pumping Station in SW ¼, sec. 33, T57N, R33W; stratigraphic height above base of Freda, about 150′; thickness sampled 400′; strike N27°E, dip 25°W. |
| KA2 | 286.0 | + 2.5 | |
| KA3 | 275.0 | −12.5 | |
| KA4 | 273.5 | −20.0 | |
| KA5 | 278.5 | + 4.0 | |

## Table V (cont'd)

| Sample | D | I | Description of locality |
|---|---|---|---|
| KA6 | 289.5 | − 1.0 | From outcrops in ditch alongside Tamarack Pumping Station road, about ½ mile from station and near boundary between sec. 33, T57N, R33W and sec. 4, T56N, R33W; at base of Freda sandstone; thickness sampled 100′; strike N20°E, dip 25°W. |
| KA7 | 266.0 | −14.0 | |
| KA8 | 265.0 | −13.0 | |
| KA9 | 267.5 | −13.0 | |
| KA10 | 275.0 | +20.0 | |
| KA11 | 287.0 | +26.0 | |
| KA12A | 288.0 | +16.5 | |
| KA12B | 289.5 | +14.0 | |
| KA13 | 311.5 | +15.0 | |
| KA14 | 293.5 | +21.0 | From road-cut on Allouez-Tamarack Pumping Station road at middle of boundary between sec. 27 and sec. 34, T57N, R33W; stratigraphic distance below base of Freda about 500′; thickness sampled 10′; strike N12°E, dip 24°W. |
| KA15 | 280.5 | − 3.0 | |
| KB1 | 288.5 | + 3.0 | From bed of Boston Creek below dam in SW ¼, sec. 2, T55N, R34W, stratigraphic height above base of Freda 2, 300′; thickness sampled 110′, strike N24°E; dip 18°W. |
| KB2 | 282.0 | + 4.0 | |
| KB3 | 285.5 | − 8.0 | |
| KB4 | 281.5 | −22.0 | |
| KB5 | 282.5 | +13.5 | |
| KB6 | 280.5 | −13.5 | |
| KB7 | 285.0 | −25.5 | |
| KB8 | 279.0 | − 0.5 | |
| KB9 | 278.0 | +19.5 | At contact of Nonesuch shale and Freda sandstone exposed along Boston Creek in NW ¼, sec. 12, T55N, R34W; thickness sampled 230′; strike N30°E, dip 26°W. |
| KB10 | 285.5 | +16.0 | |
| KB10B | 288.0 | +18.5 | |
| KB10C | 289.0 | +16.5 | |
| KB11 | 298.0 | +15.5 | |
| KB12 | 293.5 | +21.0 | |
| KB13 | 290.0 | +17.5 | |
| KB14 | 262.0 | + 9.0 | |
| KC1 | 279.0 | −14.5 | From sandstone exposed along Muggun Creek in SW ¼, sec. 17, T56N, R33W, just above contact between Freda and Nonesuch; thickness sampled 90′; strike N15°E, dip 24°W. |
| KC2 | 277.0 | − 6.0 | |
| KC3 | 284.0 | − 8.5 | |
| KC4 | 290.0 | − 6.0 | |
| KC5 | 284.5 | − 3.5 | |
| KC6 | 280.0 | −13.5 | |
| KD1 | 287.0 | − 4.0 | From outcrop along shore of Lake Superior in SW ¼, NE ¼, sec. 33, T57N, R33W; stratigraphic height above base of Freda 750′; thickness sampled 100′; strike N24°E, dip 22°W. |
| KD1B | 295.0 | − 7.5 | |
| KD2B | 272.0 | + 4.0 | |
| KD3B | 278.0 | +22.0 | |
| KE1 | 278.5 | − 5.0 | From just below dam at Rock Ridge, Mich.; stratigraphic height above base of Freda about 6,500′; thickness sampled 7′, virtually flat lying. |
| KE2 | 288.0 | − 6.0 | |
| KE3 | 290.5 | − 8.0 | |
| KE3B | 284.5 | − 4.5 | |
| KE4 | 311.0 | − 8.0 | |
| KE4B | 290.5 | −13.0 | |
| KE4C | 280.0 | −10.0 | |
| KE5 | 289.0 | −11.5 | |

Table V (cont'd)

| Sample | D | I | Description of locality |
|---|---|---|---|
| KF1 | 292.0 | − 6.0 | From sandstones exposed in bluffs on either side of harbor at Bayside, Mich.; stratigraphic height above base of Freda about 6,700′; thickness sampled 90′, virtually flat lying. |
| KF3 | 286.0 | − 6.5 | |
| KF4 | 286.5 | −10.5 | |
| KF5 | 298.0 | − 7.0 | |
| KF6 | 289.0 | − 4.5 | |
| KF7 | 282.0 | + 8.5 | |
| KF8 | 285.0 | 0.0 | |
| KF9 | 285.0 | − 0.5 | |
| KF9B | 277.0 | 0.0 | |
| KN5 | 277.5 | − 5.5 | From base of Nonesuch shale in SE ¼, sec. 27, T57N, R33W.; thickness sampled 5′, strike N32°E, dip 28°W. |
| KN6 | 281.5 | − 4.0 | |
| KV1 | 290.0 | −18.0 | From sandstones exposed along Montreal River in Wisconsin in SW ¼, sec. 17 and SE ¼, sec. 18, T47N, R1E; stratigraphic height above base of Freda 2,000′; thickness sampled 2,000′; strike N45°E, dip 75°W. |
| KV2 | 287.0 | −23.0 | |
| KV3 | 278.5 | −14.5 | |
| KV4 | 289.5 | + 7.5 | From exposures along lower part of Montreal River below dam in SW ¼, sec. 7, T47N, R1E; stratigraphic height above base of Freda 9,000′; thickness sampled 900′, strike N45°E, dip 75°W. |
| KV5 | 293.0 | − 3.5 | |
| KV6 | 317.5 | − 0.5 | |
| KV7 | 272.0 | +18.5 | |

The magnetic stability of the Freda and Nonesuch can be demonstrated by studying the directions of magnetization of strata that have been tilted different amounts. Figure 12 shows a stereographic plot of representative samples from flat-lying, gently tilted, and steeply tilted sites, the directions being taken with respect to the horizontal and not to the bedding planes. The directions of the normals to the bedding planes are also plotted as crosses on the upper hemisphere of the stereogram. In Figure 13 the same results are plotted, but this time the directions of magnetization are taken with respect to the bedding planes. In other words they have been corrected for geological tilt. Comparison between Figures 12 and 13 shows that a much more consistent direction is obtained by measuring the directions with respect to the bedding planes. This fact shows that these sediments have not acquired significant post-tilting magnetization. Furthermore, since the geologists believe that much of the tilting took place during and immediately after Keweenawan time, the magnetic stability would date from that time and not some subsequent time.

Two samples from the Freda sandstone were heated in field free space up to their Curie point, and it was found that the remanent magnetization, as measured at room temperature, decreased in a manner similar to a stable thermo-remanent magnetization or in a 'Curie decay'. In other words, at low temperatures the intensity of magnetization stayed fairly constant and did not begin to fall off until the temperature neared the Curie point. Furthermore, the direction of magnetization within experimental error did not change at all. These experiments would indicate that no unstable isothermal remanence has been superimposed on the

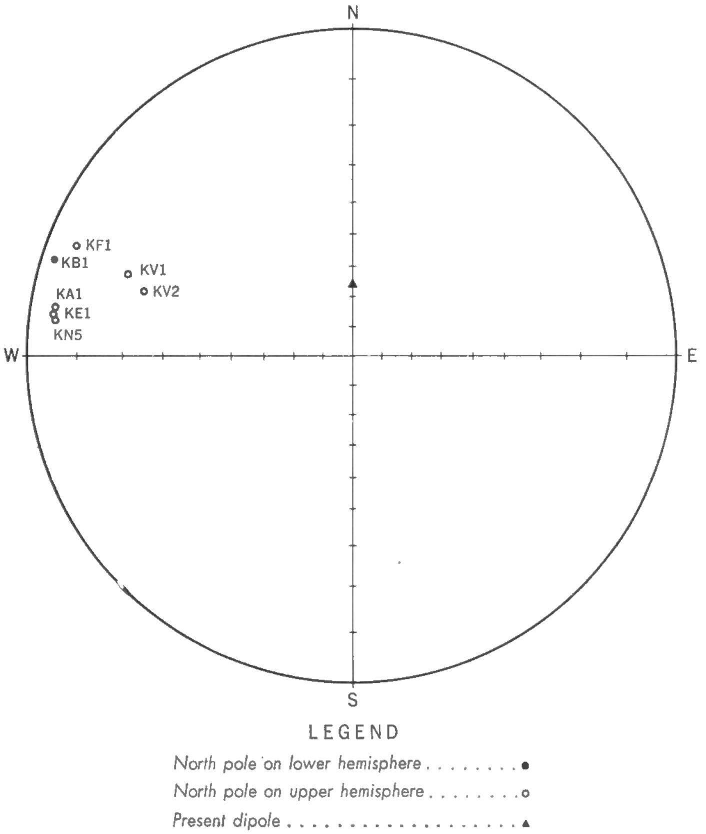


**Figure 13.** **Directions of magnetization of same samples as in Figure 12 corrected for dip.**

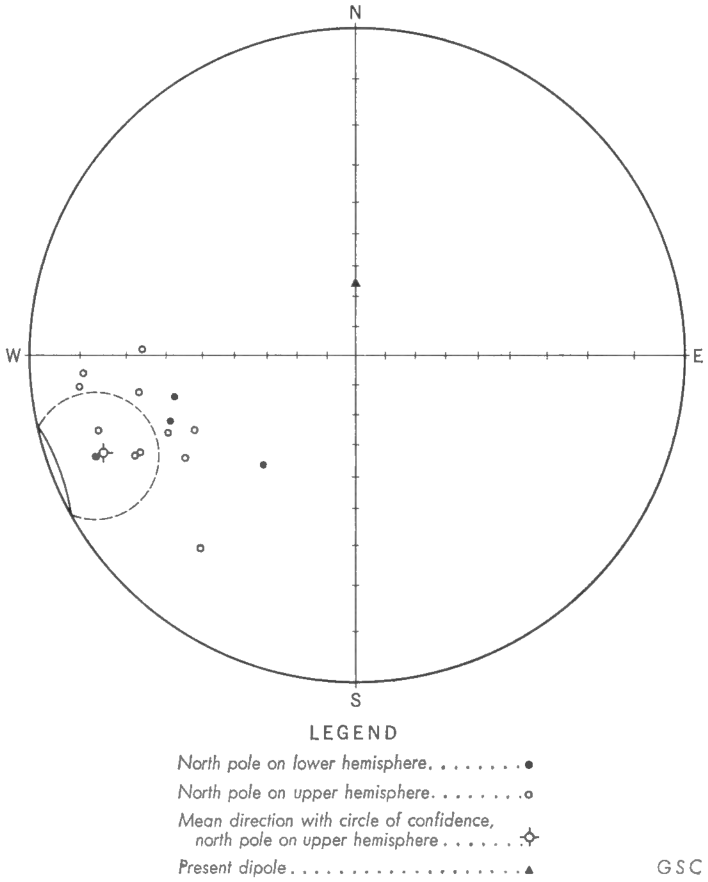


**Figure 14.** **Directions of magnetization of Jacobsville sandstone from Keweenaw Peninsula.**

original stable component. Unless the unstable component had exactly the same thermal decay properties as the stable component (and such a possibility is unlikely), one would expect the direction of magnetization to change as the specimen was heated to the Curie point. Furthermore, isothermal remanence does not have the Curie type decay with temperature.

Of course, it is conceivable that a new stable component has been superimposed on the original stable component by some such physico-chemical mechanism as an exsolution process, but even in this case, unless the thermal decay properties of the two components were identical, one would expect a change in direction of magnetization during heating. Thus there is strong evidence that the Freda is stably magnetized.

## Jacobsville Sandstone

The Jacobsville is predominantly a red to reddish brown, medium-grained sandstone characterized by lenticular bedding. There are some beds of conglomerate, particularly at the base of the formation around the margins and flanks of the old buried Precambrian hills. Although red is the predominant colour, practically all the rock is mottled with white streaks, blotches, and circular spots where the red hematite has been reduced or removed. At the centre of many of the spherical volumes of bleached rock is a black speck, which may be organic material that has reduced the hematite around it. Hamblin (1958) has recognized four main rock types in the Jacobsville sandstone: conglomerate, lenticular sandstone, massive sandstone, and red siltstone. The most important member, the lenticular sandstone, he considered to be fluvial in origin and cited the presence of clay pebbles, mudcracks, ripple-marks, and crossbedding as evidence. The massive sandstone, on the other hand, Hamblin believed to have been developed in a lacustrine environment, and there is some evidence that the massive sandstone overlies the lenticular formation. Hamblin thus tentatively concluded that the conditions during the depositions of the Jacobsville formation changed from a predominantly fluvial to lacustrine environment.

The Jacobsville sandstone is exposed in the eastern lowlands of the Keweenaw Peninsula and extends as far south as the South Range near Gogebic Lake. It extends eastwards and outcrops almost continuously along the south shore of Lake Superior as far east as Sault Ste. Marie. Hamblin (1958) has shown that the Jacobsville forms a clastic wedge that pinches out against the pre-Keweenawan rocks of the upper Michigan Peninsula and thickens to the north rapidly to several thousand feet. The source area of these sediments has been shown by Hamblin to have been to the south, in the pre-Keweenawan crystalline highlands.

The age of the Jacobsville has long been a matter of controversy because of the complete absence of fossils and the lack of exposures showing its stratigraphic relations with older and younger rocks. The only definite evidence is that the Jacobsville is younger than the Portage Lake lavas, on which it rests unconformably at the South Range, and older than the Upper Cambrian Munising sandstone, which overlies the Jacobsville with a small angular unconformity at Grand Island.

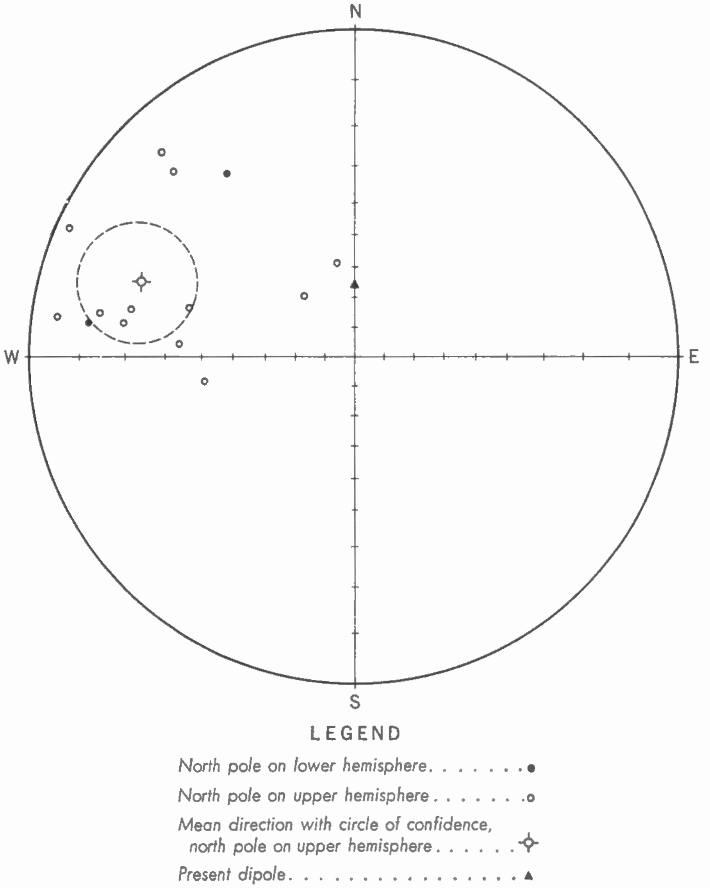


**Figure 15.** *Mean directions of Jacobsville sandstone from Sault Ste. Marie.*

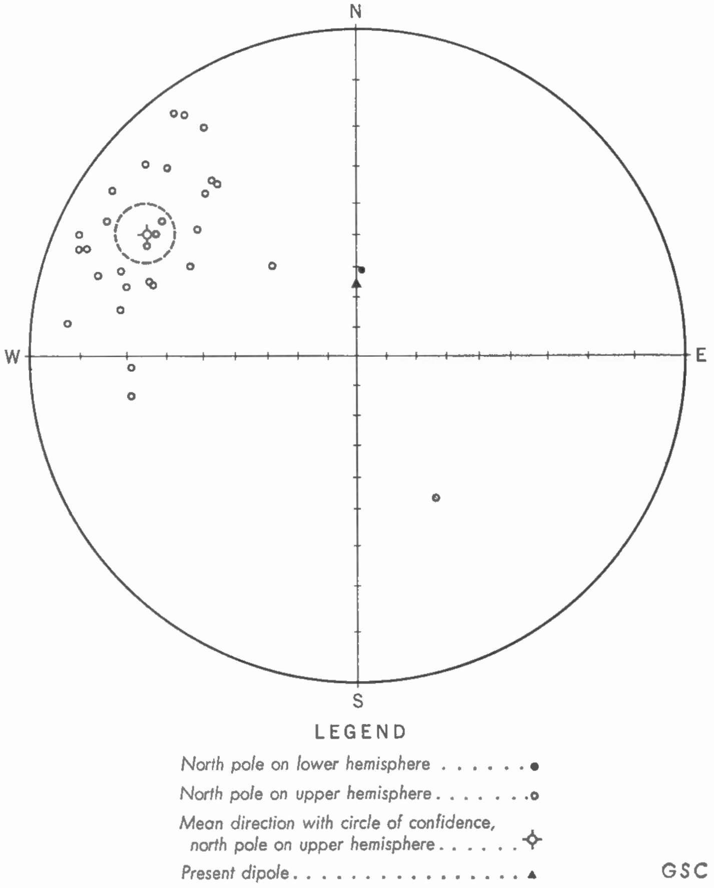


**Figure 16.** *Directions of magnetization of Eileen sandstone corrected for dip.*

Twenty-nine samples were collected and measured from the Jacobsville sandstone (*see* Table VI). As the collection sites on the Keweenaw Peninsula and at the Sault are separated by a distance such that the earth's ambient magnetic field at the two areas could have been significantly different, a direct comparison between the directions from the two areas should not be made. The results for the two areas have, therefore, been plotted on separate stereograms (Figs. 14 and 15).

### Table VI

*Results for Jacobsville Sandstone*

| Sample | D | I | Description of locality |
|---|---|---|---|
| CA1 | 260.5 | −22.0 | From red sandstone layer at mouth of Tobacco River near Gay on east shore of Keweenaw Peninsula, 15 miles east of Calumet and 22 miles NE of Jacobsville; SW ¼, sec. 20, T56N, R30W. |
| J1 | 219.5 | −15.5 | From Jacobsville quarries in SE ¼, sec. 18, T53N, R32W. |
| J4 | 239.5 | −27.5 | |
| J5 | 219.5 | +43.0 | |
| J6 | 245.5 | −32.5 | |
| J7 | 249.5 | + 9.0 | |
| J8 | 246.5 | −18.0 | |
| J9 | 248.5 | −26.5 | |
| J10 | 254.0 | −11.5 | |
| J11 | 263.5 | − 9.5 | |
| J12 | 272.0 | −23.5 | |
| J13 | 251.0 | +27.5 | |
| J14 | 266.5 | −10.5 | |
| J15 | 246.5 | −17.5 | |
| J16 | 257.5 | +30.5 | |
| LS3B | 294.0 | − 2.0 | From banks of Root River east from bridge that carries highway 17 over the river north of Sault Ste. Marie. |
| LS3P | 317.0 | − 9.0 | |
| LS4B | 351.5 | −58.0 | |
| LS4P | 325.0 | +22.0 | |
| LS5B | 259.5 | −40.5 | |
| LS5P | 286.0 | −35.0 | |
| LS6B | 279.5 | −13.5 | |
| LS6P | 320.5 | −64.0 | |
| LS7B | 281.5 | −20.0 | |
| LS7P | 274.0 | −34.5 | |
| LS8B | 316.0 | −13.5 | From south shore of Whitefish Island just below dam across St. Mary's River at Sault Ste. Marie. |
| LS8P | 278.5 | −19.0 | |
| LS9B | 277.0 | − 5.0 | |
| LS9P | 277.0 | +11.5 | |

## Eileen Sandstone

The Eileen sandstone is a red and white quartzose sandstone, that is well exposed on the South Fork of Fish Creek near Ashland, Wisconsin. There 1,800 feet of sandstone, which dip steeply to the north, is exposed in beds that vary from massive to shaly. Although the colours range from brick-red to white, red is the most prevalent, and many of the rocks exhibit much mottling and banding of red and white. The terrestrial nature of these rocks is indicated by ripple-marks, crossbedding, and the well-oxidized condition of most of the iron ores, although some irregular beds of almost pure magnetite are present. The common mineral constituents are quartz, orthoclase feldspar, mica, and the iron oxides hematite and magnetite.

During the spring of 1957 thirty samples of Eileen sandstone were collected from the steeply tilted beds exposed along the South Fork of Fish Creek (*see* Fig. 6, site 3; Table VII; Fig. 16). The rocks there strike easterly and either dip very steeply to north or have been overturned. However, the rocks show few signs of faulting, fracturing or slickensiding, or other evidence of such disturbance.

The numerical results corrected for geological dip are presented in Table VII and Figure 16. The mean direction, which is given in Table XVIII, has been calculated using all the samples except E16 and E17, which do not conform to the general grouping represented by the other samples.

Farther downstream in the NE ¼ of sec. 15, T47N, R5W steeply tilted beds of the Amnicon formation, which overlies the Eileen, are exposed in a bluff. The exposed rocks consist of coarse red and white arkoses. Six samples were collected from this locality, but the magnetizations were too scattered to be palaeomagnetically useful.

## Table VII

### *Results for Eileen Sandstone*

| Sample | D | I | Description of locality |
|---|---|---|---|
| E1 | 293.5 | − 5.5 | From site just below bridge over South Fork of Fish Creek on boundary between NE ¼ of sec. 20, T47N, R5W, and NW ¼ of sec. 21, T47N, R5W. |
| E2 | 291.5 | − 7.5 | |
| E3 | 276.0 | − 7.5 | From site about 400′ NW or upstream from E1 and E2; thickness sampled 100′. |
| E4 | 260.0 | −20.5 | |
| E5 | 320.5 | −21.5 | |
| E6 | 321.0 | −22.0 | |
| E7 | 289.0 | −24.0 | |
| E8 | 289.5 | −22.5 | |
| E9 | 304.0 | − 7.0 | |
| E10 | 299.0 | −30.5 | |
| E11 | 325.0 | − 6.5 | From bluff about 700′ upstream from site of E3-E10; thickness sampled 40′. |
| E12 | 301.5 | −19.0 | |
| E13 | 317.0 | −22.5 | |
| E14 | 315.0 | −12.5 | |
| E15 | 291.0 | − 6.5 | From localities about ½ mile downstream from Road bridge in NW ¼ of sec. 21, T47N, R5W; near the top of Eileen sandstone; thickness sampled 400′. |
| E16 | 1.0 | +61.5 | |
| E17 | 131.0 | −31.5 | |
| E18 | 281.0 | −18.0 | |
| E19 | 287.0 | −18.5 | |
| E20 | 290.0 | −16.0 | |
| E21 | 318.0 | −49.5 | |
| E22 | 267.5 | −21.5 | |
| E23 | 297.5 | −19.0 | |

*Results for Eileen Sandstone* (conc.)

| Sample | D | I | Description of locality |
|---|---|---|---|
| E24 | 298.5 | − 9.0 | About 100′ downstream from Road bridge in NW ¼ of sec. 21, T47N, R5W; thickness sampled 260′. |
| E25 | 326.0 | −10.5 | |
| E26 | 323.5 | − 5.0 | |
| E27 | 305.0 | −19.0 | |
| E28 | 308.5 | −27.0 | |
| E29 | 287.5 | −12.0 | |
| E30 | 312.5 | − 9.0 | |

## Orienta Sandstone

The Orienta sandstone is the lowest member of the Bayfield group of Thwaites. It is typically made up of brown, red, and white feldspathic sandstones that rarely show ripple-marks and crossbedding. The Orienta overlies the Amnicon formation, and Thwaites (1912) believed that the contact between the two is conformable and gradational. It underlies the Devil's Island sandstone apparently conformably, although the actual contact of the two formations is not exposed. The Devil's Island sandstone is a thin but very characteristic formation, which is distinguished by its thin bedding, by well-rounded, medium-sized quartz grains, by ripple-marks, and by its pink and white colour. It is overlain by the Chequamegon sandstone. The Orienta, particularly the upper part, closely resembles the Chequamegon, but it contains more feldspar, mica and magnetite, and this difference becomes more pronounced in the lower part of the Orienta. Due to the lack of outcrops it is very difficult to compute the thickness of the Orienta, but Thwaites makes a rough estimate of 3,000 feet.

Thirteen samples of the Orienta were collected for palaeomagnetic measurements (*see* Fig. 6, site 4), the results of which are listed in Table VIII and

**Table VIII**

*Results for Orienta Sandstone*

| Sample | D | I | Description of locality |
|---|---|---|---|
| 01 | 238.0 | + 8.0 | From shore of Lake Superior in the SE ¼, sec. 29, T50N, R9W; strata flat lying. |
| 02 | 286.0 | − 2.5 | |
| 03 | 306.0 | +62.5 | From east bank of Iron River in NW ¼, sec. 10, T49N, R9W; strata flat lying. |
| 04 | 354.5 | +57.5 | |
| 05 | 279.0 | −16.0 | |
| 06 | 291.0 | +29.0 | |
| 07 | 339.5 | +73.0 | |
| 08 | 335.0 | +65.5 | |
| 09 | 315.0 | + 2.5 | |
| 010 | 271.0 | +11.0 | From west bank of Iron River in NW ¼, sec. 10, T49N, R9W; strata flat lying. |
| 011 | 293.5 | +62.0 | |
| 013 | 241.0 | −32.5 | |
| 014 | 252.0 | −30.0 | |

presented graphically in Figure 17. This suite of rocks is unfortunately rather restricted in its geographic and stratigraphic distribution, and outcrops are too infrequent for adequate sampling. The palaeomagnetic results themselves are rather scattered. There appear to be two groups of samples. Group II, consisting of 03, 04, 07, 08 and 011, has a magnetization that differs from the present earth's field in northern Wisconsin by only 14 degrees of arc, and as the circle of confidence is 12 degrees, there is little significant difference between the two. The rest of the samples, which constitute Group I are scattered and significantly different from the present earth's field. The mean directions of Group I and Group II taken individually and together are listed in Table IX.

Table IX

*Mean Directions of Orienta*

| | Group II | Group I | Groups I and II |
|---|---|---|---|
| D | 325.5 | 272.5 | 283.5 |
| I | +66.0 | −5.0 | +25.0 |
| K | 41.8 | 6.5 | 3.1 |
| θ | 12.0 | 22.0 | 25.5 |
| N | 5 | 8 | 13 |

## Chequamegon Sandstone

The Chequamegon sandstone consists of about 1,000 feet of red and white sandstone composed predominantly of quartz grains, with thin lenticular beds of red sandy shale. This sandstone is the uppermost Keweenawan formation known and occupies the centre of the western part of the Lake Superior basin, where there appears to be a very thick sedimentary sequence of Upper Keweenawan clastic rocks.

Fifteen samples were collected from two sites in the Chequamegon (*see* Fig. 6, site 5). This collection is from too small a thickness and too restricted an area to be completely satisfactory, but the results are very consistent (*see* Table X and Fig. 18). It can be seen that the mean direction is close to that of the present earth's field, and the question arises whether the remanent magnetism of the Chequamegon is an unstable isothermal remanence. However, the mean direction, though close, is significantly different from that of the present earth's field, and this difference would indicate that these sediments have a certain amount of magnetic stability.

Further evidence on the stability of the Chequamegon is provided by heating tests in which the decay of the natural remanent magnetization with increase in

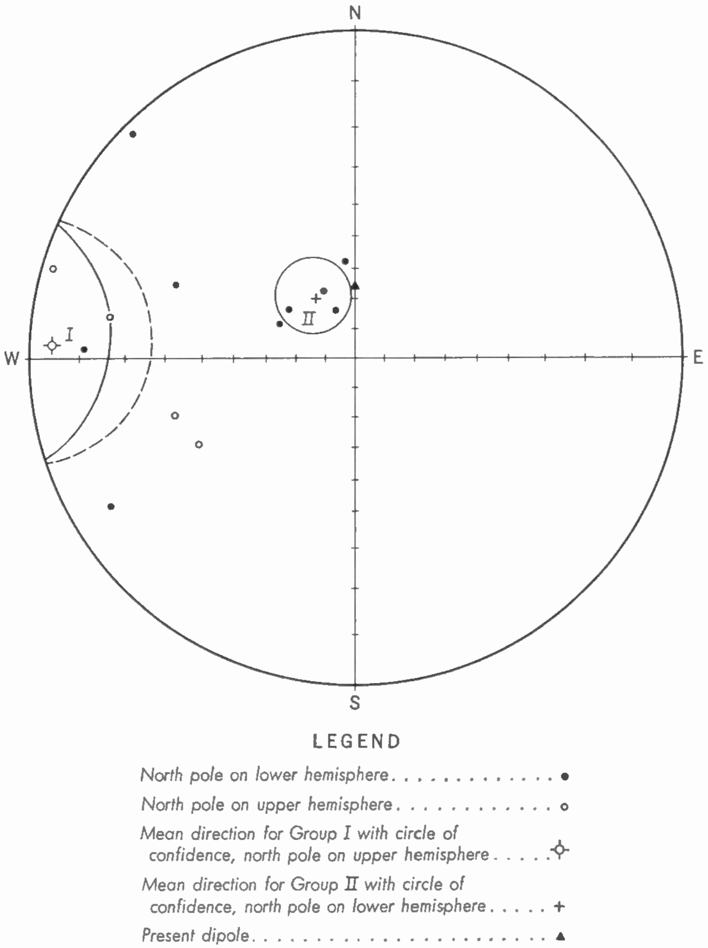


**Figure 17.** Directions of magnetization for Orienta sandstone.

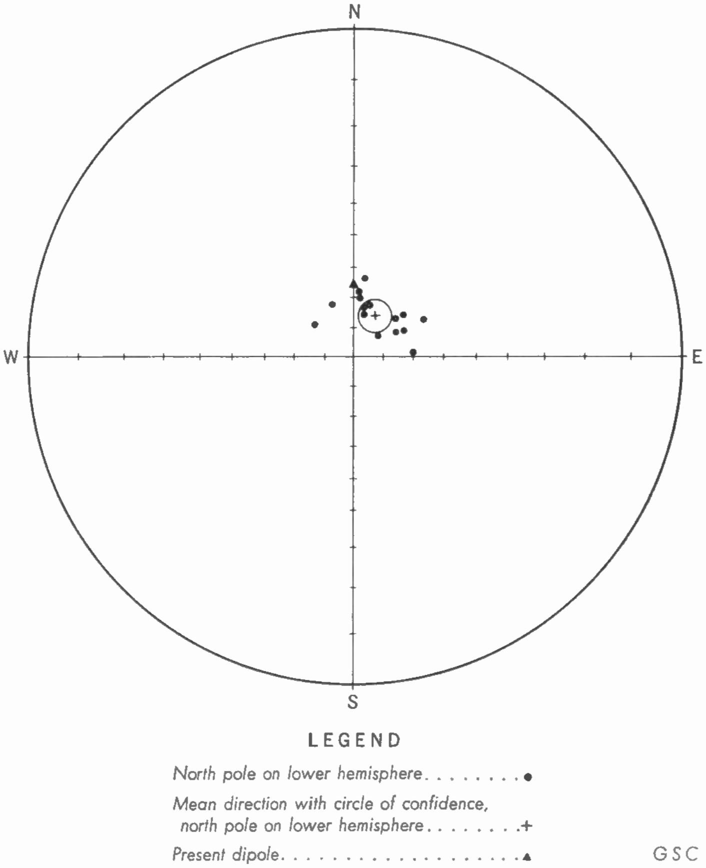


**Figure 18.** Directions of magnetization for Chequamegon sandstone.

temperature was studied. The natural remanence decreased in a 'Curie type decay' similar to stable thermoremanence. Furthermore, the direction of magnetization during these experiments remained constant.

## Table X

*Results for Chequamegon Sandstone*

| Sample | D | I | Description of locality |
|---|---|---|---|
| K1 | 15.0 | +75.0 | From quarry south of Fish Hatchery near Bayfield on shore of Lake Superior in sec. 33, T50N, R4W; strata flat lying. |
| K2 | 15.0 | +72.5 | |
| K3 | 318.0 | +73.5 | |
| K4 | 63.0 | +63.0 | |
| K5 | 14.0 | +72.5 | |
| K6 | 64.5 | +70.5 | |
| K7 | 51.0 | +67.5 | |
| K8 | 51.5 | +78.5 | |
| K9 | 88.0 | +69.5 | |
| K10 | 64.5 | +73.0 | |
| K11 | 6.0 | +68.0 | |
| K12 | 49.5 | +70.5 | |
| K13 | 9.5 | +63.0 | From lake shore just north of Fish Hatchery in sec. 23, T50N, R4W; strata flat lying. |
| K14 | 7.5 | +70.0 | |
| K15 | 338.5 | +71.0 | |

## Table XI

*Results for Duluth Gabbro*

| Sample | D | I | Description of locality |
|---|---|---|---|
| D1B | 287.0 | +34.0 | On north side of road running between secs. 15 and 22 in SW ¼, SE ¼, sec. 15, T50N, R15W. |
| D1P | 276.5 | +45.5 | |
| D2B | 292.5 | +21.0 | On west side of road running between secs. 21 and 22, SE ¼, NE ¼, sec. 21, T50N, R15W. |
| D2P | 276.5 | +43.5 | |
| D3B | 309.5 | +37.5 | On south side of road running between secs. 21 and 28, NW ¼, NE ¼, sec. 28, T50N, R15W. |
| D3P | 309.5 | + 6.5 | |
| D4B | 289.5 | +28.5 | On west side of road running between NE¼ and NW ¼ of sec. 33, NE ¼, NW ¼, sec. 33, T50N, R15W. |
| D4P | 293.5 | +37.5 | |
| D5B | 306.0 | +46.0 | On east side of road running between secs. 27 and 28, SW ¼, sec. 27, T50N, R15W. |
| D5P | 268.0 | +36.5 | |
| D6B | 292.5 | +28.5 | On south side of road running between secs. 14 and 23, NW ¼, NE ¼, sec. 23, T50N, R15W. |
| D6P | 292.5 | +22.5 | |
| D7B | 285.0 | +33.0 | On east side of road running between secs. 25 and 26, NW ¼, SW ¼, sec. 25, T50N, R15W. |
| D7P | 259.5 | +29.0 | |
| D8B | 296.0 | +48.5 | On Minnesota Route 53, SE ¼, SW ¼, sec. 18, T50N, R14W. |
| D8P | 293.0 | +36.5 | |
| D9B | 238.0 | + 2.0 | From hill overlooking Bay of St. Louis in NW ¼, SW ¼, sec. 32, T50N, R14W. |
| D9P | 280.0 | +34.5 | |

## Duluth Gabbro

During the spring of 1958 eighteen samples were collected from the Duluth gabbro near the city of Duluth (*see* Fig. 6, site 6, and Table XI). All the outcrops are to be found on the maps accompanying Minnesota Geological Survey Bulletin 33. The magnetic measurements are tabulated in Table XI and are presented graphically in Figure 19. The mean direction and Fisher's statistics have been computed and are presented in Table XVIII.

## Sibley Sediments

Twenty-four samples were collected from the Sibley series at two main localities, Sibley Peninsula and the valley of Black Sturgeon River (*see* Fig. 6, site 9). In the type locality, Sibley Peninsula, Tanton (1931) divided the series into a six-fold classification which is shown below.

| | | *Feet* |
|---|---|---|
| (F) | Grey grit and quartz sandstone interbedded and intermixed with red mudstone | 40+ |
| (E) | Red and purple mudstone | 50–350 |
| (D) | Thinly interbedded grey chert and limestone | 2 |
| (C) | White quartz sandstone | 40 |
| (B) | Pink limestone, red mudstone, and white quartz sandstone | 60 |
| (A) | Basal conglomerate | 0–8 |

The basal conglomerate, which is present in most localities where the base of the Sibley is exposed, lies in a horizontal position at the southern end of Sibley Peninsula on top of the flat-lying Animikie. There is no obvious unconformity between the two formations, but north and east of Loon Lake the Sibley sediments overlap the Animikie and rest on early Precambrian rocks. Thus, although the two formations appear structurally conformable, there is probably a disconformity between them.

The C member forms striking cliffs along the western side of Sibley Peninsula from Thunder Cape to north of Pass Lake.

The E member is the most extensive and important member of the Sibley series and consists principally of poorly bedded red and purple mudstone. Some of the beds have bleached areas, where the hematite has apparently been removed by leaching or reduction, but most of the beds exhibit various shades of red, bright brick-red being the most common.

To the northeast of Sibley Peninsula are the lower Keweenawan sediments of the Nipigon basin, which can be divided into four units, basal conglomerate, sandstone, shale, and dolomite. These sediments are mostly characterized by a bright red colour and bleached areas where the hematite has been removed or reduced. Although the Sibley cannot be traced laterally to the Nipigon sediments, there seems little doubt that they are equivalent. They are very similar in lithology,

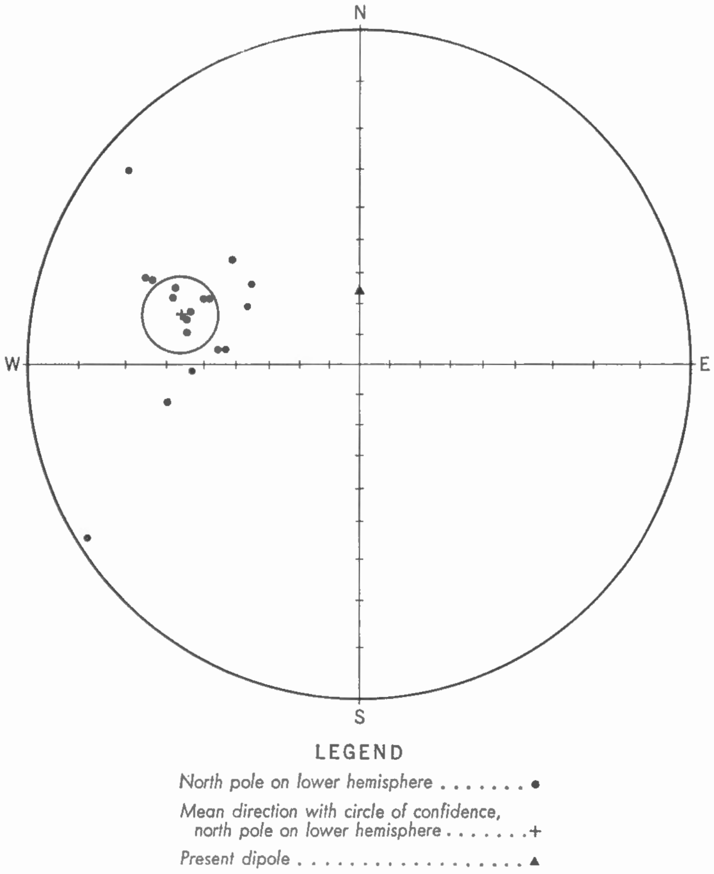


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Figure 19. **Directions of magnetization for Duluth gabbro.**

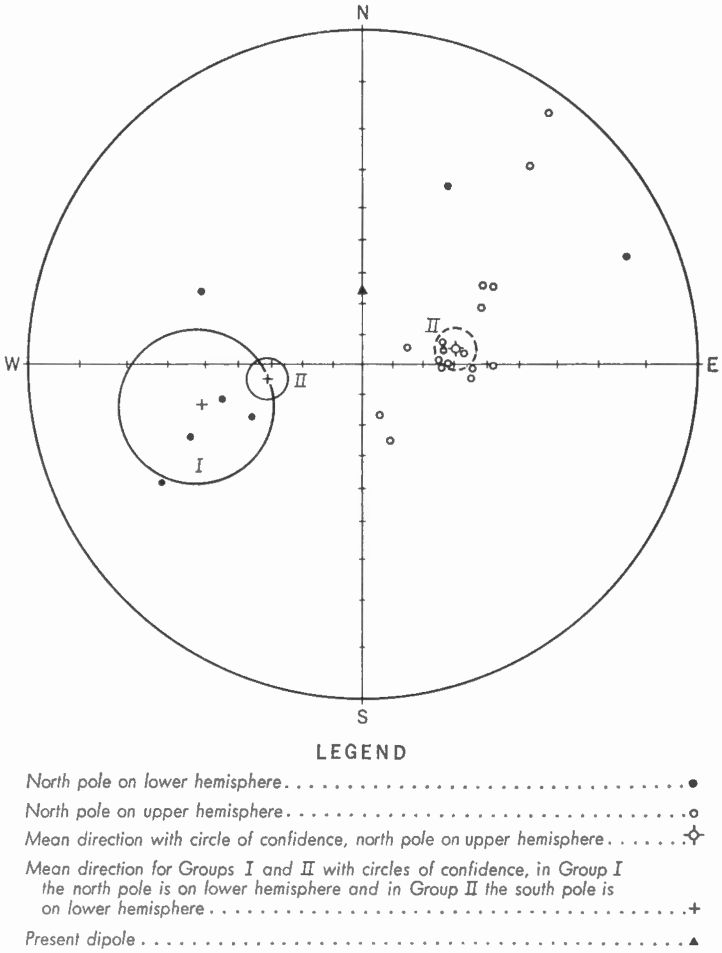

Figure 20. **Directions of magnetization of Sibley sediments.**

except that the dolomite is limited to the Nipigon basin where it forms the upper part of the sediments. It is possible that the Sibley may lie unexposed beneath the water separating Sibley Peninsula and Edward Island. In the Nipigon area the dolomites are the most important member of the series because of their extensive distribution and thickness. They are generally fine grained and range in colour from dark red, almost brown, to green and greenish grey.

The numerical results of the measurements are given in Table XII, and Figure 20. The results fall into three main groups. The first group (I) consists of rocks that have a west-southwestward declination and moderate dips downward. The second group (II) has an east-northeastward declination with upward dips, which are greater in magnitude than those of group I. The third group has directions in the northeast quadrant with shallow positive and negative dips. Two samples N26B and N26P were apparently baked by an overlying diabase sheet and therefore do not represent the field at the time of their sedimentation. Groups I and II are approximately reversed to each other, and in order to test whether there was any significant difference between their axes of magnetization, the mean direction and circle of confidence for each group was calculated and plotted on Figure 20. The mean direction for Group II was actually reversed so that it could be com-

## Table XII

*Results for Sibley Series*

| Sample | D | I | Description of locality |
|---|---|---|---|
| N2P | 93.0 | −62.5 | From flat lying, brick-red dolomites exposed for about a mile in shallow gorge cut by Black Sturgeon River just before it enters Nonwatin Lake. |
| N3B | 97.0 | −53.5 | |
| N3P | 92.0 | −53.0 | |
| N4B | 91.0 | −47.0 | |
| N5B | 75.5 | −62.0 | |
| N5P | 81.0 | −62.0 | |
| N6B | 71.5 | −73.5 | |
| N6P | 65.0 | −47.0 | |
| N7B | 87.5 | −63.5 | |
| N7P | 84.0 | −55.5 | |
| N8B | 91.8 | −61.0 | |
| N8P | 57.5 | −43.5 | |
| N26B | 160.0 | −72.0 | From red mudstone exposed along CPR tracks at Red Rock; probably baked by overlying diabase. |
| N26P | 159.5 | −62.5 | |
| SS1 | 294.5 | +34.5 | Red mudstones exposed near Red Sandstone Lake on Sibley Peninsula; all samples within 40′ of one another. |
| SS2 | 26.0 | +29.0 | |
| SS3 | 60.0 | −41.0 | |
| SS4 | 37.0 | − 4.0 | |
| SS5 | 68.0 | + 9.5 | |
| SS6 | 244.5 | +50.0 | From small quarry exposed on Pearl road northeast of Silver Islet landing. |
| SS7 | 239.0 | +20.5 | |
| SS8 | 255.5 | +43.5 | |
| SS9 | 247.0 | +31.5 | |
| SS10 | 40.5 | −14.5 | From cliff on highway 17 southwest of Nipigon. |

pared with the mean direction of Group I. It can be seen that these two axes of magnetization are not significantly different from each other. The mean direction for all the samples was then computed and plotted on the upper hemisphere. In all these calculations the specimens of the third group were left out on the assumption that they represented samples which were only partly stable. N26B and N26P were also excluded for obvious reasons.

## Logan Sills and Dykes, Group I

During the spring of 1957 and of 1958, one hundred and nineteen samples were collected from the Logan sills and dykes, in an area from the International Boundary at Pigeon River to east and northeast of Nipigon, Ontario (*see* Fig. 6, sites 7 and 8). These samples have been divided for discussion into two groups on the basis of their remanent moment of magnetization. The first group, comprising eighty-seven samples, has a steep upward or reversed magnetization, whereas the second group of thirty-two samples has a magnetization with a westward declination and moderate downward or normal dip, similiar to the results for the Keweenawan lavas and Duluth gabbro. Most of the rocks of the first type occur in extensive sills, but a few are from small dykes. On the other hand all the samples of the second group are from prominent northeasterly striking dykes although there are a few related dykes that strike northwesterly.

It is worth pointing out here that samples P8B and P8P, which belong to the first group, are from what appears to be part of the diabase sill about Jarvis Bay. This probable sill is cut by the dyke from which samples P9B and P9P, which have been placed in the second group, were taken. Furthermore, samples P7B and P7P of the second group are from a very prominent northwesterly striking dyke which Tanton was able to trace and map on the ground and which is very conspicuous on aerial photographs. The study of these photographs indicates that this dyke cuts the Logan sills to the northwest of Pigeon Point. As all the major sills between Port Arthur and Pigeon River appear to have group I magnetization as the first group, this evidence would indicate that the dyke from which samples P7P and P7B were taken was intruded after the sills.

The rocks and results of group I have been subdivided into two subgroups, A and B, Group IA was collected from the sills about Nipigon, and group IB was from the sills near Port Arthur. The mean results for each group have been computed separately and together and are given in Tables XIV and XVIII.

As the sills are almost flat lying, the measurements were made with respect to the horizontal. Some of the results quoted were obtained after A.C. demagnetization, details of which are described on page 11.

Specimens SN30 to SN35 are all from the lower thin sill on Mount Mackay and have scattered directions of magnetizations; even within one sample the directions for two cubes may be different (*see* Table XV). It is possible that this scattering may be due to the superimposition of a random magnetic component

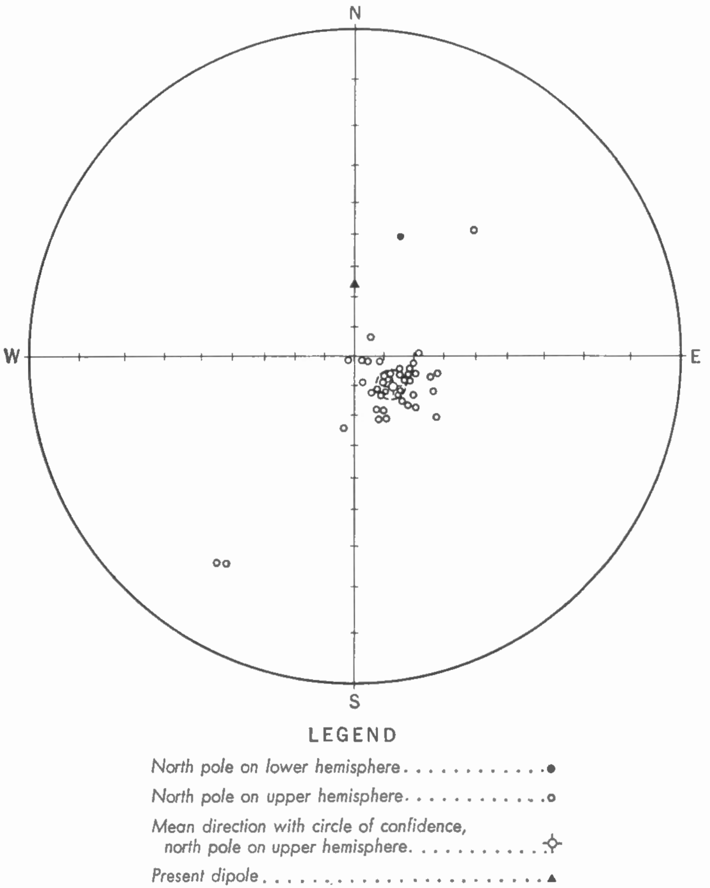

**Figure 21.** **Directions of magnetization of group IA, Logan diabase sills and dykes.**

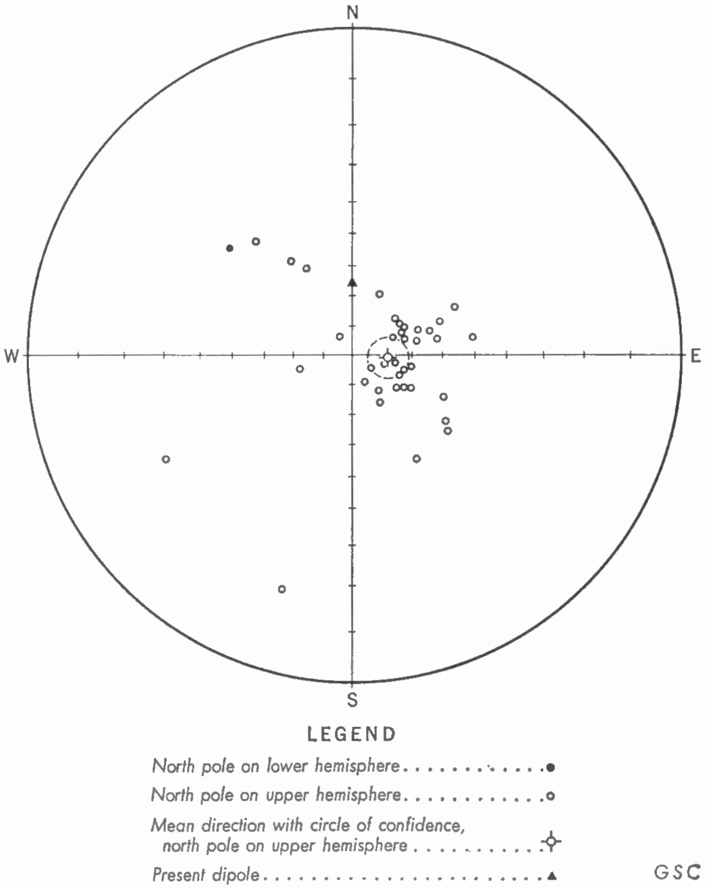

**Figure 22.** **Directions of magnetization of group IB, Logan diabase sills and dykes.**

onto the original permanent moment. The Mount Mackay sill because of its elevation may have been subject to the effect of lightning, which can impart disordered magnetizations to a rock.

## Table XIII

*Results for Logan Sills and Dykes, Group I*

| Sample | D | I | Description of locality |
|---|---|---|---|
| N1B | 157.0 | −68.0 | About 30′ below dam on Black Sturgeon River, on north side of river near Camp No. 8 of Great Lakes Paper Co. (Group A). |
| N1P | 116.5 | −70.0 | |
| N9B | 43.5 | −34.5 | From top of diabase sill just south of Ozone Station east of Nipigon (Group A). |
| N9P | 21.0 | +48.5 | |
| N10B | 115.5 | −72.0 | From lower contact of same sill as N9 but about ½ mile farther south (Group A). |
| N10P | 134.0 | −69.5 | |
| N11B | 146.5 | −74.5 | |
| N11P | 158.5 | −70.5 | |
| N12B | 134.0 | −70.0 | From sill exposed in a scarp just southwest and south of Nipigon (Group A). |
| N12P | 159.5 | −67.0 | |
| N13B | 133.5 | −66.0 | |
| N13P | 163.5 | −80.5 | |
| N14B | 124.0 | −66.5 | |
| N14P | 109.5 | −71.0 | |
| N15B | 130.5 | −75.0 | |
| N15P | 125.0 | −75.0 | |
| N16B | 115.5 | −76.5 | From sill exposed near bridge by which highway 17 crosses river 2 miles southwest of Ozone Station (Group A). |
| N16P | 130.0 | −63.0 | |
| N17B | 95.5 | −70.0 | |
| N17P | 114.5 | −61.0 | |
| N18B | 211.5 | −17.0 | |
| N18P | 213.5 | −16.0 | |
| N19B | 126.5 | −56.5 | From scarp along highway 11 just south of Orient Bay and opposite cabins on Reflection Lake (Group A). |
| N19P | 235.0 | −87.5 | |
| N20B | 126.5 | −73.0 | |
| N20P | 121.5 | −78.0 | |
| N21B | 101.5 | −61.5 | From same sill, 1 mile south of N20 (Group A). |
| N21P | 103.0 | −70.0 | |
| N22B | 105.0 | −69.0 | |
| N22P | 88.0 | −68.5 | |
| N23B | 37.5 | −81.5 | From sill exposed in road-cut on Pine Portage road just north of Lake Maria (Group A). |
| N23P | 104.0 | −63.5 | |
| N24B | 142.0 | −74.5 | From diabase exposed in road-cut on Pine Portage road about 3 miles south of Cameron Falls (Group A). |
| N24P | 125.5 | −71.5 | |
| N25B | 101.5 | −81.0 | From diabase sill about 2 miles northwest of Nipigon (Group A). |
| N25P | 107.5 | −74.0 | |
| N27B | 128.0 | −72.0 | From cliff exposure above CPR and CNR tracks at Red Rock. N27's and N28's from sediment immediately below diabase sill; sediment apparently altered by heat of sill from red sandstone to a bluish purple, banded, well-indurated rock; zone of alteration extends about 40′ below sill. N29's from lower part of sill. N30's from a 3′-wide dyke that cuts both sediments and sill (Group A). |
| N27P | 155.5 | −76.5 | |
| N28B | 152.5 | −69.0 | |
| N28P | 148.0 | −76.0 | |
| N29B | 116.5 | −87.5 | |
| N29P | 111.5 | −85.0 | |
| N30B | 117.0 | −73.0 | |
| N30P | 189.0 | −65.5 | |

*Results for Logan Sills and Dykes, Group I* (conc.)

| Sample | D | I | Description of locality |
|---|---|---|---|
| N31PA | 145.5 | −74.5 | From diabase exposed on eastern scarp of Sleeping Giant sill. N31PA and N31PB from dyke that cuts Animikie and Sibley sediments and Sleeping Giant sill; other samples from sill (Group B). |
| N31PB | 116.0 | −56.5 | |
| N32B | 321.0 | −83.0 | |
| N32P | 151.5 | −71.0 | |
| N33B | 83.0 | −50 0 | |
| N33P | 149.0 | −49.0 | |
| N34B | 310.0 | +37.0 | |
| N40B | 107.5 | −71.5 | From dyke exposed on logging road leading from Silver Islet road to Sleeping Giant (Group B). |
| N40P | 66.0 | −53.0 | |
| P8B | 326.0 | −63.0 | From what appears to be diabase sill at Jarvis Bay; may be part of sills about Jarvis Bay that are cut by dyke of samples P9B and P9P (Group B). |
| P8P | 121.5 | −68.5 | |
| P11B | 330.5 | −57.5 | From what appears to be a diabase dyke near Cloud Bay resort (Group B). |
| P11P | 196.5 | −16.5 | |
| SN5 | 22.5 | −68.0 | From diabase sill which is exposed along highway 17 on hill just northeast of bridge which crosses the Current River. The Current reservoir dam is built on this sill (Group B). |
| SN6 | 67.0 | −72.5 | |
| SN7 | 64.5 | −71.5 | |
| SN8 | 103.5 | −70.0 | |
| SN9 | 69.5 | −58.5 | |
| SN10 | 52.5 | −71.5 | |
| SN11 | 74.0 | −71.5 | |
| SN12 | 78.5 | −68.5 | |
| SN13 | 58.0 | −72.0 | |
| SN14 | 114.0 | −72.5 | |
| SN15 | 68.0 | −75.5 | |
| SN16 | 241.0 | −23.5 | From diabase in NW ¼, lot 12, con. V, Pardee tp.; at centre of large negative aeromagnetic anomaly; impossible to determine from field evidence if dyke or sill, but size of anomaly suggests that body has great depth and is therefore a transgressive intrusion (Group B). |
| SN17 | 111.0 | −75.0 | |
| SN18 | 126.5 | −52.0 | |
| SN19 | 318.5 | −41.5 | |
| SN20 | 129.5 | −49.0 | |
| SN21 | 129.0 | −71.0 | |
| SN22 | 129.5 | −82.0 | |
| SN23 | 156.5 | −79.0 | |
| SN24 | 122.5 | −69.5 | |
| SN25 | 254.0 | −71.5 | |
| SN26 | 74.0 | −63.0 | From diabase that forms Horne Falls on Pigeon River (Group B). |
| SN27 | 105.5 | −77.5 | |
| SN28 | 70.5 | −66.5 | |
| SN29 | 81.0 | −61.0 | |
| SN30 | *See* Table No. 15 | | From lower diabase sill of Mount Mackay near Fort William, exposed in parking lot (Group B). |
| SN31 | | | |
| SN32 | | | |
| SN33 | | | |
| SN34 | | | |
| SN35 | | | |

Whatever the mechanism by which these rocks acquired their present scattered magnetizations, in the two cubes from a single rock sample the difference in direction may be presumed to be the result of a difference in the ratio between the strengths of the superimposed and original components. Then by the principles

## Table XIV

*Results of Logan Sills*

| —— | Group IA | Group IB | Group I |
|---|---|---|---|
| D | 128.5 | 99.0 | 117.0 |
| I | −74.0 | −77.5 | −76.0 |
| Θ | 5.5 | 7.0 | 4.5 |
| K | 16.8 | 10.8 | 13.2 |
| N | 43 | 37 | 80 |

of vector addition the directions of magnetizations of the two cubes should determine the plane that contains both the original magnetization and the superimposed component.

In Figure 23 the directions of magnetization of the cubes of SN30 to SN35 have been plotted on a stereogram. Where the directions of magnetization of the cubes of the same sample are sufficiently different, a great circle has been drawn through the poles that represents the plane determined by the two vectors. If the superimposed component is different for each rock sample, then the intersection of these great circles should give the direction of the original stable

## Table XV

*Results for Mount Mackay Sill*

| Specimen | D | I |
|---|---|---|
| SN30.1 | 151.5 | − 3.0 |
| .2 | 140.0 | −31.0 |
| [1]SN31 | 152.5 | − 2.0 |
| SN32.1 | 110.0 | −51.5 |
| .2 | 128.5 | −50.0 |
| SN33.1 | 301.5 | −16.5 |
| .2 | 297.0 | +24.5 |
| [1]SN34 | 107.0 | −42.0 |
| SN35.1 | 199.5 | −34.0 |
| .2 | 197.5 | −34.0 |

[1]only one cube cut.

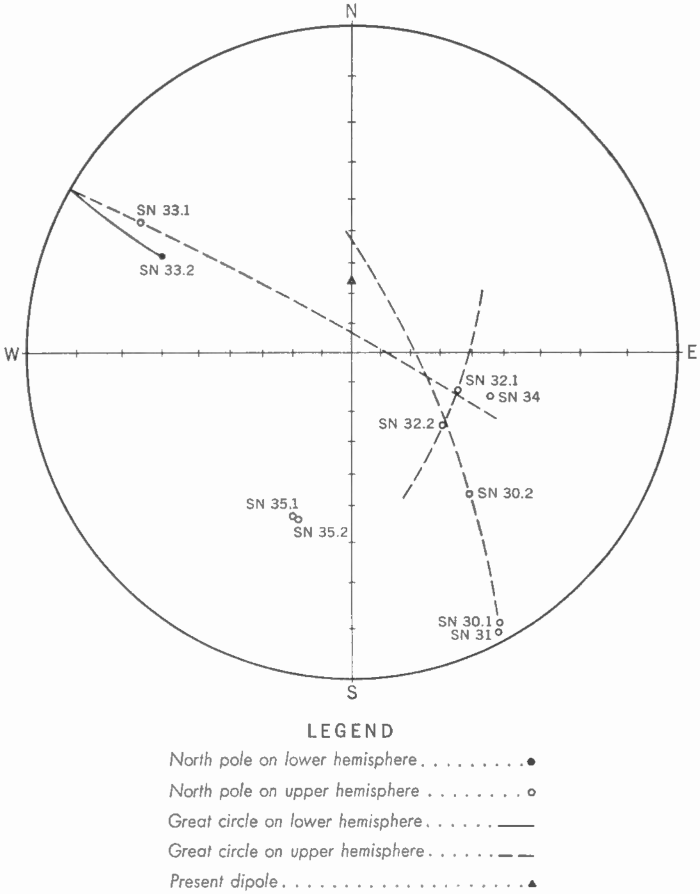

**Figure 23.** **Directions of magnetization of lower Mount Mackay sill.**

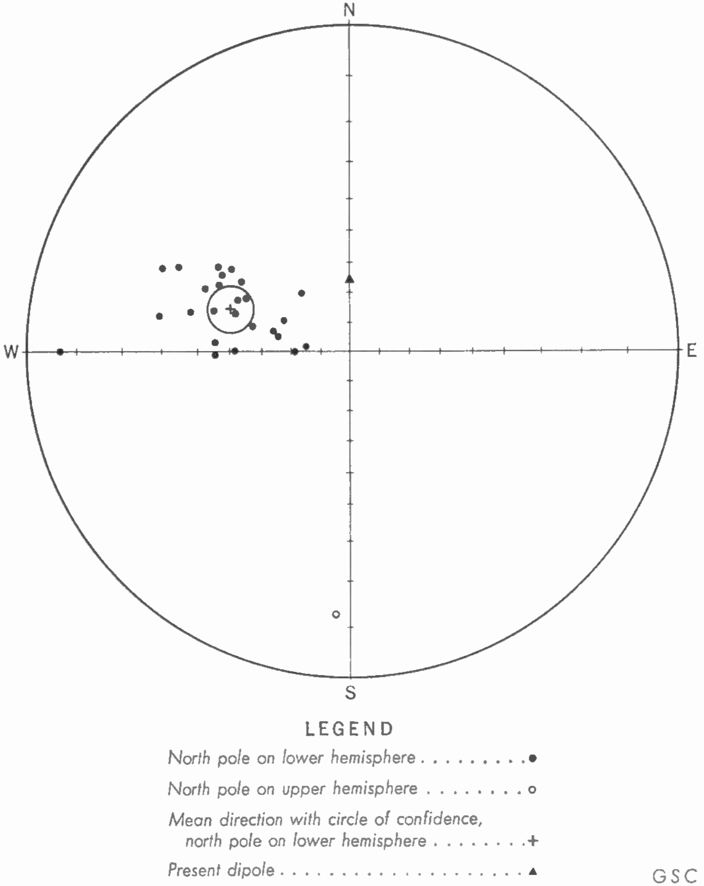

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**Figure 24.** **Directions of magnetization of group II, Logan diabase dykes.**

component. It can be seen from the figure that the intersection does in fact occur very close to the mean direction determined from the uniformly magnetized diabase sills.

These samples were subjected to magnetic washing in the hope that the superimposed component was sufficiently soft to be removed by A.C. demagnetization, but it was found to be too stable.

## Logan Sills and Dykes, Group II

The thirty-two samples that comprise group II of the Logan intrusions, have been grouped on the basis of the similarity of their remanent moments of magnetization (*see* Table XVI, Fig. 24). In Figure 24 the results from all samples are graphically plotted except for the N35's, N36's, and N37's. These are omitted because they are weakly magnetized and the directions of magnetization widely divergent.

The exceptional samples are all from one dyke. This dyke forms Silver Islet and Pyritic Island and has been the site of important mineralization, especially on Silver Islet itself where considerable silver was mined. It is possible that the hydrothermal solutions that mineralized parts of this dyke were responsible for the destruction of its original permanent moment. Perhaps palaeomagnetic measurements might be used to discover those dykes that have been affected by hydrothermal solutions and that, therefore, may carry mineral deposits.

## Discussion of Results for Logan Sills and Dykes

The mean directions of magnetization for the two groups of samples from the Logan sills and dykes are given in Table XVIII (Nos. 4 and 8), with the parameters of Fisher's statistics. In group I samples SN30-35 and N34B and in group II samples N35's, N36's, N37's and P4B were excluded from the calculation because of internal inconsistency.

In group I the mean remanent moment of magnetization has a steep upward dip or is, in other words, a reverse magnetization. The first natural question is whether this reversed magnetization is due to some physico-chemical mechanism such as proposed by Néel (1955) and Graham (1953) or whether it is due to a reversed magnetic field at the time the rocks were formed. Two good arguments can be made for the case of reversal of the earth's magnetic field. Firstly, all the samples are reversely polarized over an area more than 100 miles long. Most of the physico-chemical mechanism that have been proposed to explain reversely magnetized rocks are sensitive to such factors as composition and rate of cooling. Over such a large area of intrusion these factors would be expected to vary widely and, therefore, in some rocks the magnetization would be normal and in others reversed. In these rocks it is uniformly reversed. Secondly, four samples N27B, N27P, N28B and N28P are from sediments that were baked by an overlying sill, and they are also reversely magnetized. It is improbable that both the sill and the baked sedi-

## Table XVI

### *Results for Logan Sills and Dykes, Group II*

| Sample | D | I | Description of locality |
|---|---|---|---|
| N35B<br>N35P | 20.0<br>143.5 | +80.0<br>+10.5 | From northeast tip of Silver Islet near old silver mine workings. |
| N36B<br>N36P<br>N37B<br>N37P | 9.0<br>340.0<br>323.0<br>235.0 | +61.5<br>+64.0<br>+56.5<br>+66.5 | From south shore of Pyritic Island; this dyke is probably a southwestward extension of Silver Islet dyke. |
| N38B<br>N38P | 270.5<br>268.0 | +51.0<br>+71.0 | From south shore of Shangoina Island about 50′ west of suspected fault and mineralized zone. |
| N39B<br>N39P | 296.0<br>296.0 | +42.0<br>+51.0 | From southwestern tip of Burnt Island dyke. |
| P1B<br>P1P | 294.5<br>286.0 | +65.0<br>+43.0 | From dyke that forms the 90′-high Pigeon Falls on Pigeon River. |
| P2B<br>P2P | 268.5<br>273.0 | +45.0<br>+45.0 | From dyke ¼ mile downstream from Pigeon Falls. |
| P3B<br>P3P | 281.5<br>284.0 | +64.5<br>+63.0 | From dyke ¾ mile north of Pigeon Falls. |
| P4B<br>P4P | 181.5<br>319.5 | −13.0<br>+65.0 | From dyke that crosses Scott Highway in SE ¼, lot 6, con. VIII, Crooks tp. |
| P5B<br>P5P | 274.0<br>288.0 | +75.5<br>+50.0 | From dyke northeast of Lenore Lake, Pardee tp. |
| P6B<br>P6P | 293.0<br>283.5 | +49.0<br>+55.0 | From dyke just south of Crystal Lake, Pardee tp. |
| P7B<br>P7P | 301.5<br>270.0 | +39.0<br>+ 7.0 | From northwesterly striking dyke that crosses Scott Highway in Mining Location 57B, Crooks tp. |
| P9B<br>P9P | 280.0<br>295.5 | +28.5<br>+29.5 | From dyke ½ mile northwest of Jarvis Bay. |
| P10B<br>P10P | 304.0<br>300.0 | +42.5<br>+41.0 | From dyke exposed about 2 miles east of Cloud Bay P.O. on road to Jarvis Bay. |
| SN1<br>SN2 | 293.0<br>293.5 | +38.5<br>+26.0 | From dyke at Silver Islet Landing. |
| SN3<br>SN4 | 301.5<br>283.5 | +47.0<br>+36.5 | From dyke just north of above dyke. |

ment had the capacity for self-reversal, because they are entirely different in composition and structure. Furthermore, it is worth noting that the mean direction of group I is significantly different from the anti-parallel direction of group II. This shows that the axis of the magnetic vector, without regard to sign, of group I is different from that of group II.

Because the direction of magnetization for the sills intruded into the lower Keweenawan sediments near Nipigon cannot be distinguished from the similar sills intruded into the Animikie near Port Arthur, the two groups of sills are probably very close to one another in age. Thus, with respect to most of the sills in this region, there appears to have been one major period of intrusion. Furthermore, samples P8B and P8P of group I are from a diabase mass that appears to be part of the diabase sill at Jarvis Bay. As this sill is cut by the dyke from which samples P9B and P9P of group II were collected, the magnetization of group I represents a direction of the earth's magnetic field that antedates that of group II. This conclusion is supported by the fact that the dyke from which samples P7B and P7P of group II were collected cuts extensive and prominent sills that with little doubt should be correlated with the sills of group I. The relationships of these palaeomagnetic results with other results from the Lake Superior region are discussed more fully on page 56.

## Sault Ste. Marie Lavas

During the spring of 1958 a total of twenty-two samples of Keweenawan lava were collected: twelve were from the lavas exposed around Alona Bay and ten were from Mamainse Point (*see* Fig. 6, sites 10 and 11). In each of these two localities, two samples of interbedded sediments were also collected. At Alona Bay, two medium-grained red sandstone dykes were found, from each of which two samples were collected. Four samples were collected from a gently tilted sandstone overlying the lavas.

Brief descriptions of the localities from which the samples were collected and the magnetic measurements corrected for geological dip are given in Table XVII and Figures 25 and 26.

The results from the Alona Bay lavas are fairly consistent, except for samples KL4B and KL4P, which are from a mineralized outcrop. It is possible that if this mineralization occurred during middle or upper Keweenawan time these two samples acquired their magnetization then rather than at the time of extrusion. This extrusion, the palaeomagnetic evidence for which is discussed later, may well have taken place in lower rather than in middle Keweenawan time.

The mean directions for both groups of lavas and interbedded sediments have been computed and are given in Table XVIII, No. 1. KL4B and KL4P have been omitted from the calculation.

The two mean directions are strikingly different. Not only are the Alona Bay lavas reversely magnetized from the Mamainse Point group, but they have a significantly different axis of magnetization. As both groups strike north-south and dip to the west, and as the Alona Bay lavas are northeast of Mamainse Point, it is possible that the lavas at Alona Bay are older than those at Mamainse Point. This difference in age may explain the difference in magnetization. These points, however, are discussed more fully on page 59.

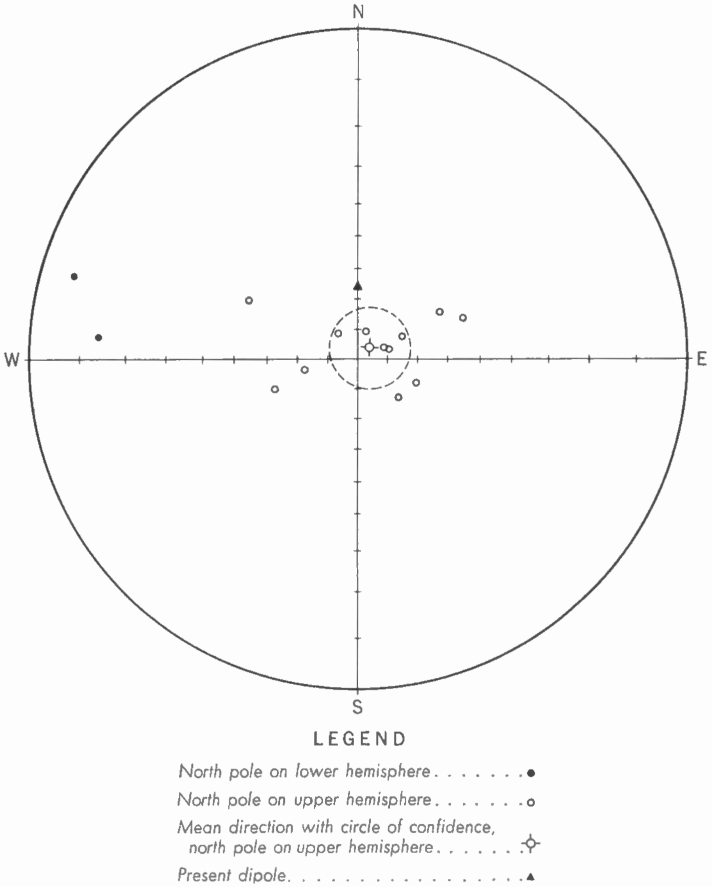

Figure 25. **Directions of magnetization of Alona Bay lavas.**

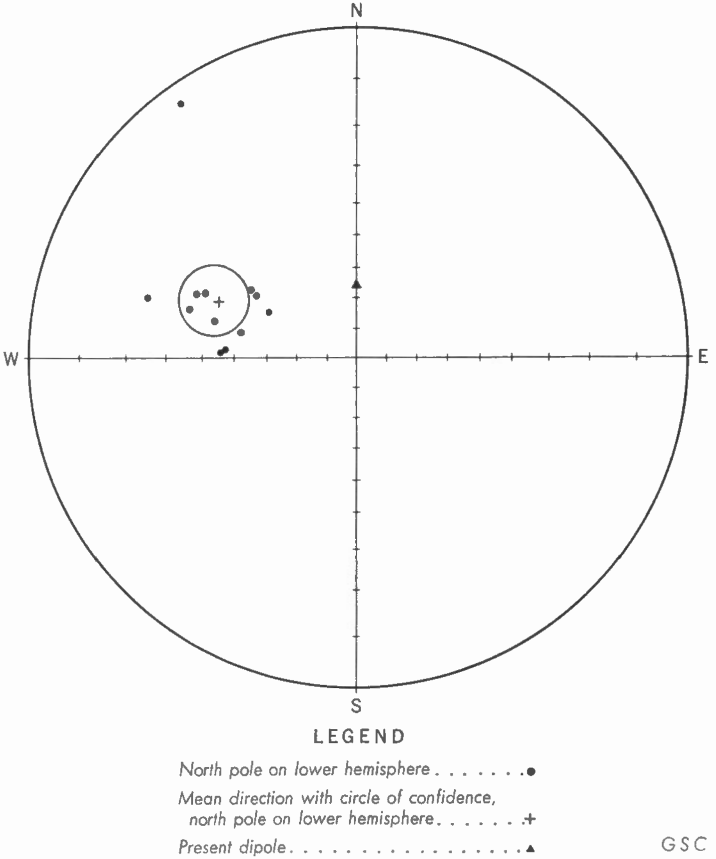

Figure 26. **Directions of magnetization of Mamainse Point lavas.**

The results for the overlying sandstones and sandstone dykes, corrected for dip, are given in Table XVII, and Figures 27 and 28, and the mean results in Table XVIII.

## Table XVII

*Results for Lavas and Sediments North of Sault Ste. Marie*

| Sample | D | I | Description of locality |
|---|---|---|---|
| KL1B<br>KL1P | 256.0<br>322.5 | −71.5<br>−79.5 | From shore of Alona Bay near mouth of creek just below bend in highway 17; strike N 3°E, dip 24°W. |
| KL2B<br>KL2P | 249.0<br>298.5 | −60.5<br>−50.0 | From shore of Alona Bay about ½ mile east of Point aux Mines; strike N 6°W, dip 30°W. |
| KL3B<br>KL3P | 113.0<br>133.0 | −68.0<br>−71.0 | From point of land about 1/8 mile northeast of contact of lavas with underlying granite gneiss; strike N 13°E, dip 38°W. |
| KL4B<br>KL4P | 286.0<br>274.5 | + 7.0<br>+14.5 | From shore of Alona Bay about 300 yards southwest of KL3B and KL3P; strike N 13°E, dip 38°W. |
| KL5B<br>KL5P | 71.0<br>73.5 | −80.0<br>−79.0 | About 100 yards northeast of KL4B and KL4P; strike N 13°E, dip 38°W. |
| KL6B<br>KL6P | 17.0<br>63.0 | −80.0<br>−73.0 | From shore of Alona Bay near old copper workings; strike N 6°W, dip 30° W. |
| KS1B<br>KS1P | 61.0<br>69.0 | −58.0<br>−52.0 | From interbedded fine-grained brick-red siltstone near KL2B and KL2P; strike N 6°W, dip 30°W. |
| LS1B<br>LS1P | 322.0<br>299.0 | +31.5<br>+14.5 | From beach on Alona Bay northeast of lavas and just below highway 17; strike N 81°E, dip 20°N. |
| LS2B<br>LS2P | 314.0<br>290.5 | +21.5<br>+ 8.5 | From beach on Alona Bay about ¼ mile southwest of LS1; strike N 63°E, dip 17°NW. |
| LSD1B<br>LSD1P | 238.0<br>233.0 | +54.5<br>+76.0 | From sandstone dyke cutting lavas and calcite veins near samples KL2B and KL2P; strike and dip of lavas N 6°W, 30°W. |
| LSD2B<br>LSD2P | 272.0<br>208.5 | +58.5<br>+53.0 | From sandstone dyke about 200 yards northeast of LSD1B and LSD1P, strike and dip of lavas N 6°W, 30°W. |
| KL7B<br>KL7P | 284.5<br>293.0 | +42.0<br>+37.0 | From small peninsula southwest of Hibbard Bay on Mamainse Point in southwest corner of Sand Bay location of Montreal Mining Co.; strike N 36°W, dip 20°SW. |
| KL8B<br>KL8P | 272.0<br>273.0 | +45.5<br>+46.5 | From lava about 150 yards northeast of KL7B and KL7P; strike N 36°W, dip 20°SW. |
| KL9B<br>KL9P | 282.5<br>302.5 | +50.5<br>+48.5 | From roadside-cut on highway 17 near side-road leading to Mamainse Harbour settlement; strike N 3°W, dip 45°W. |
| KL10B<br>KL10P | 325.0<br>291.5 | + 3.5<br>+35.0 | From roadside-cut on highway 17, 3/4 mile northeast of Mamainse Harbour; strike N 6°W, dip 40°W. |
| KL11B<br>KL11P | 286.0<br>285.5 | +34.5<br>+23.0 | From roadside-cut on highway 17 near entrance to U.S. Army Artillery range; strike N 5°W, dip 30°W. |
| KS2B<br>KS2P | 301.5<br>297.0 | +51.0<br>+56.5 | From interbedded, red-brick, fine-grained sandstone 200 yards southwest of KL7B and KL7P; strike N 36°W, dip 20°SW. |

## *Chapter V*

# DISCUSSION OF RESULTS

In the previous sections the palaeomagnetic data have been presented as collections of measurements on individual rock samples. For each defined rock-unit a mean direction can be calculated, and Fisher's statistics provide a method of estimating the circle of confidence for each mean direction. If two rock-units, for each of which the mean palaeomagnetic direction is known, are from the same locality, then it is possible to compare their magnetizations directly. If the rocks have been collected from localities well separated from each other, then such a direct comparison cannot be made. There are, however, two methods by which these mean directions can be compared. The first is to compute from each mean direction its equivalent dipole magnetic pole positions; the pole positions so computed can then be compared. In the second method, after the pole positions have been computed for each unit, the fields that would be produced at some specified locality by these dipoles can be calculated: these fields can then be compared directly with one another. The mean palaeomagnetic direction for each unit studied is presented in Table XVIII. The positions of one of the two poles associated with the equivalent dipole and the field direction that this dipole would produce at the point 45°N, 90°W have also been listed. Included in these results are the palaeomagnetic measurements of Graham (1953) on Keweenawan dykes of Baraga county, Michigan.

In Figure 29, on which the pole positions have been plotted with their ovals of confidence, the known geological sequences have been indicated by drawing arrows from the older poles to the younger ones. It can easily be seen that the pole has wandered relatively to North America during Keweenawan times in a general northeast-to-southwest path across the central Pacific. An exception is the lowermost Keweenawan, represented by the Sibley.

## Lower Keweenawan

As the palaeomagnetism of the Logan diabase sills (4)[1] near Port Arthur and those near Nipigon are virtually identical, there can be little doubt that all these sills are contemporaneous. They intrude both the Animikie and lower Keweenawan strata, and are therefore definitely younger than either formation.

The palaeomagnetic results from the Sibley series indicate that there is a significant time gap between their deposition and the intrusion of the Logan sills.

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[1] In this chapter numbers in parentheses refer to formation numbers in Table XVIII and Figure 29.

# Table XVIII

## *Complete Palaeomagnetic Results for Keweenawan of Lake Superior*

| *Formation | Collecting site | | Mean direction at collecting site | | | | | Mean direction at 45°N 90°W | | | | | Pole position | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | Lat. | Long. | D | I | Θ | K | N | D′ | I′ | Θ | K′ | N′ | Lat. | Long. | α | β |
| 18. Chequamegon | 47.0N | 91.0W | 30.5 | +74.0 | 5.5 | 46.5 | 15 | 28.5 | +73.5 | 5.5 | 46.5 | 15 | 68.0S | 131.5E | 9.0 | 9.0 |
| 17. Mean Jacobsville | — | — | — | — | — | — | — | 266.5 | −12.0 | — | — | — | 6.5S | 179.0E | — | — |
| 16. Jacobsville (Keweenaw). | 47.0N | 88.5W | 249.5 | −10.5 | 12.5 | 10.6 | 15 | 248.5 | − 7.0 | 12.5 | 10.6 | 15 | 17.5S | 166.0W | 6.5 | 13.0 |
| 15. Jacobsville (Sault) | 47.0N | 84.5W | 289.0 | −21.0 | 14.0 | 10.5 | 12 | 285.0 | −15.5 | 14.0 | 10.5 | 12 | 5.0N | 164.0E | 7.0 | 15.0 |
| 14. Orienta I | 47.0N | 91.0W | 272.5 | − 5.0 | 22.0 | 6.5 | 8 | 273.0 | − 6.0 | 22.0 | 6.5 | 8 | 0.0 | 176.0E | 11.0 | 22.0 |
| 13. Eileen | 47.0N | 91.0W | 300.0 | −17.5 | 6.5 | 18.0 | 28 | 300.5 | −20.5 | 6.5 | 18.0 | 28 | 13.0N | 150.5E | 3.5 | 7.0 |
| 12. Freda and Nonesuch | 47.0N | 88.5W | 284.5 | − 0.5 | 3.5 | 25.8 | 68 | 283.5 | 0.0 | 3.5 | 25.8 | 68 | 9.0N | 170.0E | 2.0 | 3.5 |
| 11. Sault Freda (?) | 47.0N | 84.5W | 305.5 | +19.5 | 17.0 | 24.8 | 4 | 303.0 | +23.0 | 17.0 | 24.8 | 4 | 31.0N | 164.5E | 8.5 | 17.5 |
| 10. Copper Harbor | 47.0N | 88.5W | 293.5 | +32.5 | 6.5 | 20.2 | 25 | 293.5 | +33.0 | 6.5 | 20.2 | 25 | 29.0N | 176.5E | 4.5 | 7.0 |
| 9. Duluth gabbro | 47.0N | 92.0W | 285.5 | +33.0 | 8.5 | 17.0 | 18 | 287.5 | +30.0 | 8.5 | 17.0 | 18 | 23.5N | 179.5E | 6.0 | 9.5 |
| 8. Logan diabase II | 48.0N | 89.5W | 288.5 | +47.5 | 6.5 | 21.1 | 25 | 290.0 | +47.0 | 6.5 | 21.1 | 25 | 33.0N | 171.5W | 5.5 | 8.0 |
| 7. Mamainse Point lavas | 47.0N | 84.5W | 292.0 | +40.5 | 10.0 | 19.2 | 12 | 289.5 | +44.5 | 10.0 | 19.2 | 12 | 31.5N | 173.5W | 9.0 | 12.0 |
| 6. Portage Lake | 47.0N | 88.5W | 282.5 | +44.0 | 4.0 | 47.2 | 31 | 282.5 | +45.0 | 4.0 | 47.2 | 31 | 26.5N | 168.5W | 3.0 | 5.0 |
| 5. Sandstone dykes | 47.0N | 84.5W | 242.5 | +65.5 | 25.5 | 13.7 | 4 | 237.5 | +68.5 | 25.5 | 13.7 | 4 | 19.0N | 123.0W | 25.0 | 39.0 |
| 4. Logan diabase I | 48.5N | 89.0W | 117.0 | −76.0 | 4.5 | 13.2 | 80 | 123.0 | −75.0 | 4.5 | 13.2 | 80 | 53.5N | 130.0W | 8.0 | 8.0 |
| 3. Alona Bay lavas | 47.0N | 84.5W | 49.5 | −84.5 | 13.5 | 11.6 | 12 | 33.5 | −86.0 | 13.5 | 11.6 | 12 | 39.0N | 95.0W | 25.0 | 25.0 |
| 2. Baraga Co. dykes | 47.0N | 88.5W | 82.0 | −86.0 | — | — | — | 90.0 | −87.0 | — | — | — | 45.0N | 99.5W | — | — |
| 1. Sibley | 48.5N | 89.0W | 78.0 | −51.5 | 7.5 | 23.1 | 18 | 79.0 | −52.5 | 7.5 | 23.1 | 18 | 16.0N | 149.0W | 7.5 | 9.5 |

* Numbers refer to pole positions on Figure 29.
α and β represent the semi-axes of the oval of confidence.

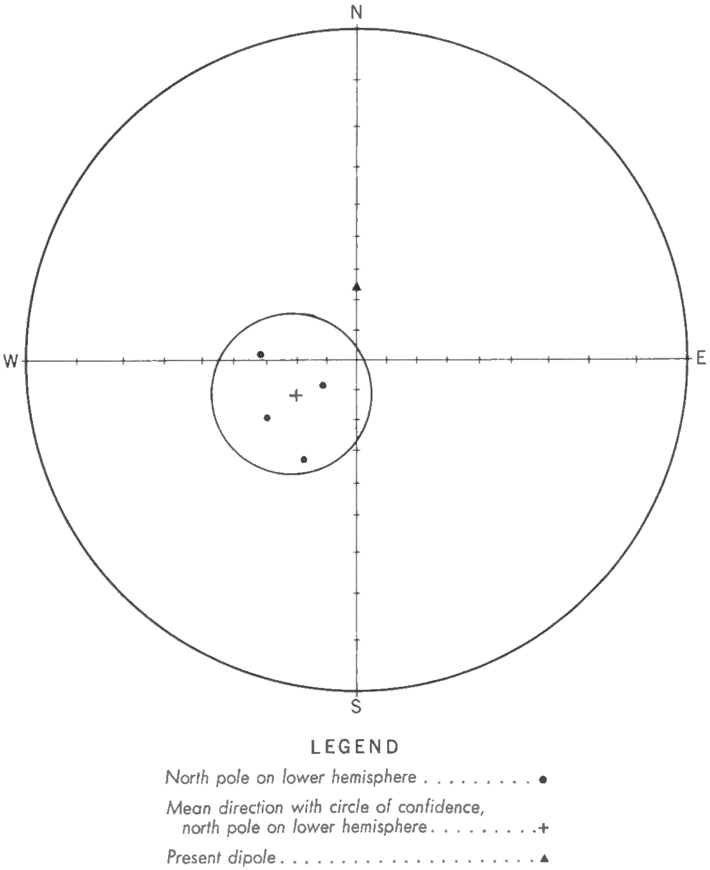

**Figure 27.** **Directions of magnetization of sandstone dykes cutting Alona Bay lavas.**

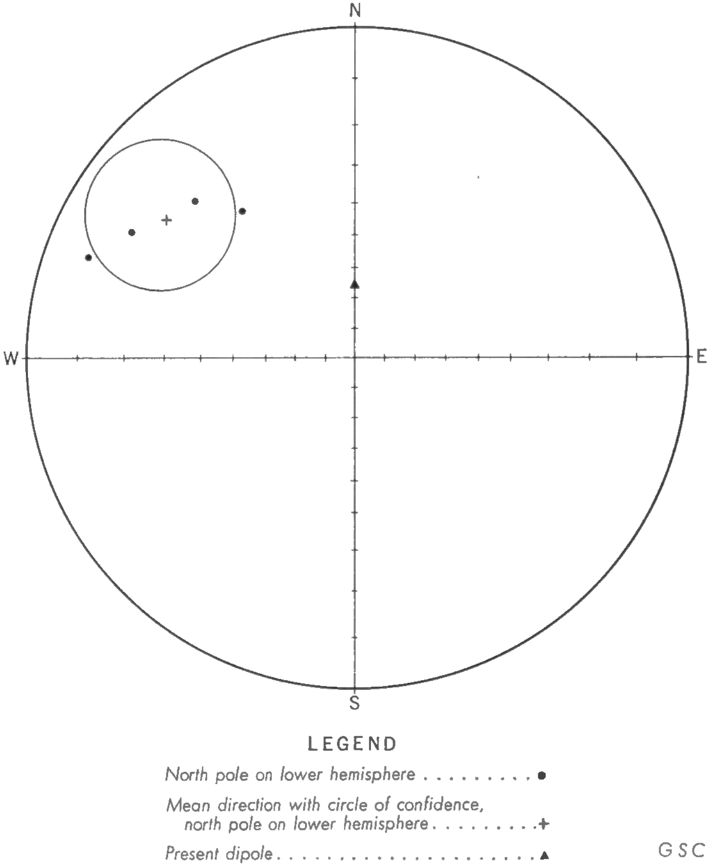

**Figure 28.** **Directions of magnetization of sandstones overlying Alona Bay lavas.**

Farquhar and Russell (1957) from their isotope studies of the anomalous leads found in veins cutting the lower Keweenawan sediments and Logan sills have put an upper age of 1,700±30 million years for the lead mineralization. Unfortunately this age tells nothing about the age of the sediments and diabase. The Logan sills near the Canadian-American border have been metamorphosed by the heating action of the Duluth gabbro, which has been recently dated by Goldich, *et al.* (1957) as about 1,120 million years old. Certain limits can be put on the age of the Animikie, which underlies the Sibley series. Goldich, *et al.* have found that potassium-argon ages show that the Thomson formation, which is usually correlated with the Animikie, was metamorphosed 1,690 million years ago. The Cuyuna iron-formation and the iron-formations of northern Wisconsin and Michigan, which are generally considered to be Animikie, also were probably metamorphosed 1,700 million years ago. Goldich, *et al.*, have calculated from potassium-argon ratios for feldspars from granite at Babbitt, Minnesota, an age of 2,100 million years, assuming 35 per cent argon loss from the feldspars. As this granite is overlain by the Animikie sediments, the age of the Animikie would lie somewhere between 1,700 and 2,100 million years.

Graham (1953) has reported on the palaeomagnetism of certain dykes from Baraga county, northern Michigan. Some of these dykes, which occur in an east-trending swarm, were found to have consistent magnetizations with steep upward dips. Graham's measurements are reported in Table XVIII. These dykes were considered to be Keweenawan intrusions and possibly feeders for the lavas, because of their petrological similarity. However, the computed pole position is very different from that of the Portage Lake lavas, but does fall close to the circle of confidence of the Logan sills (the circle of confidence for the dykes was not computed because Graham's results were not tabulated). These dykes intrude Animikie-type rock and are overlain by flat-lying Jacobsville sandstone. By using the palaeomagnetic data these dykes would be correlated with the Logan sills, and this correlation is consistent with the geological evidence.

North of Sault Ste. Marie, the Alona Bay lavas have a magnetization that gives a pole position not significantly different from that of the Logan sills. These lavas, on the basis of lithology, have usually been correlated with the Mamainse Point and Portage Lake lavas, but their magnetization is quite different. Rather it suggests strongly that they should be correlated with the Logan sills; there is no significant difference in their pole positions.

It should be pointed out that these lavas are very restricted in distribution and the figures for their magnetization may not be sufficiently representative to be significant. If, however, these palaeomagnetic results are accepted, then there is a suggestion of a period of igneous activity, both intrusive and extrusive, in the Lake Superior basin in lower Keweenawan time. Tanton (1931) found pebbles of red porphyry in the basal conglomerate of the Osler series of Ontario. He suggested that these pebbles may have come from lavas that were extruded prior to the main outpouring of Keweenawan lavas.

The absolute age of lower Keweenawan time can be fixed within certain limits. It is certainly older than the Duluth gabbro, which is 1,120 million years, and younger than the Animikie, whose age probably lies between 1,700 and 2,100 million years. On Sibley Peninsula the contact between the Animikie and overlying Sibley shows no angular discordance, but regional studies, which show that the basal member of the Sibley lies on all members of the Animikie and even on the pre-Animikie basement, would indicate that there is disconformity between the Animikie and Sibley. The magnitude of the time break represented by this disconformity cannot, however, be estimated. If these lower Keweenawan formations can truly be correlated with one another, then some additional evidence can be obtained concerning absolute ages from the Alona Bay area. There, certain basic dykes have been dated (Cumming, *et al.,* 1955) as older than 1,200 million years by the lead isotope ratio method. The galenas used were found in shear zones along the margins of the dykes where lead and uranium mineralization has taken place. These dykes were classified by Nuffield (1956) as lower Keweenawan, because they are not found to cut the Alona Bay lavas, which he considered to be middle Keweenawan. If these shear zones were formed and mineralized shortly after the emplacement of these dykes, the Alona Bay lavas would then be no older than 1,200 million years. However, the mineralization might have occurred much later, even after the extrusion of the Alona Bay lavas, and there is therefore no conclusive evidence about the absolute age of the lavas. There is, however, a good chance that they are less than 1,200 million years old.

The Baraga county dykes were intruded into metamorphosed Animikie rocks. The dykes themselves are fresh and unaltered with chilled edges, and stand perpendicularly. They were therefore probably intruded well after the deformation and metamorphism of the Animikie country rocks. This metamorphism in northern Michigan has been dated by a number of workers by both the potassium-argon and strontium-rubidium methods. Wasserburg, *et al.* (1956) have reported a potassium-argon age of 1,610 million years for muscovite from a pegmatite in a post-Animikie granite of the Felch area.

James (1958) reported that unpublished determinations made by Aldrich, Wetherill, and Davis by the strontium-rubidium method on micas from several samples of metamorphosed Animikie strata and from post-Animikie—pre-Keweenawan granitic rock yielded values about 1,400 million years. Therefore, it is probable that the Baraga county dykes are less than 1,400 million years old. All in all, the lower Keweenawan is probably between 1,120 and 1,400 million years old and the Animikie between 1,700 and 2,100 million years old.

## Middle Keweenawan

Goldich, *et al.* (1957) have determined the ages of a number of formations from northeastern Minnesota by the potassium-argon method. They found an age of 1,200 million years for granite samples that had been recrystallized by the heating action of the Duluth gabbro. This age was confirmed by a potassium-argon age determination on biotite from the gabbro itself which gave a figure of 1,120 million years. There seems little doubt, therefore, that the Duluth gabbro was intruded between 1,120 and 1,200 million years ago.

Because the Duluth gabbro intrudes the Keweenawan lavas, the latter must be the older formation. Geologists have long considered the gabbro and the lavas as manifestations of the same igneous activity, because they are similar in petrology and structure, but there has never been any conclusive geological evidence in support of this hypothesis. However, the palaeomagnetic results for the lavas and gabbro are very similar. The pole computed for the Duluth gabbro (Fig. 29, No. 9) lies 12 degrees in longitude west of and 3 degrees in latitude south of the Portage Lake pole (Fig. 29, No. 6). On the basis of Fisher's statistics this difference is just significant and may reflect a small time gap between the extrusion of the lavas and the intrusion of the gabbro. It is worth noting that such a small movement of the pole is consistent with the overall picture of a general path of polar wandering from the northeast to the southwest. However, in the larger picture the Duluth gabbro and Portage Lake lavas must be very close to one another in age.

The pole position of the Copper Harbor conglomerate is significantly different from the Portage Lake lavas but not from the Duluth gabbro. Therefore, it is possible that the gabbro and conglomerate are almost contemporaneous with one another and represent the last stage of Keweenawan igneous activity. There is a certain amount of indirect geological evidence indicating that the Copper Harbor conglomerate is slightly younger than the gabbro. Lane is quoted on page 382 of Monograph 52 (Van Hise and Leith, 1911) as saying that some of the Copper Harbor includes numerous pebbles of red rock and gabbro and that, if he is correct in his identification of the materials, there is evidence of an erosion of sufficient magnitude during middle Keweenawan time to expose the plutonic rocks at the surface. There is also considerable evidence for a period of acidic intrusion between the extrusion of the Portage Lake lavas and the sedimentation of the Copper Harbor conglomerate. On the Keweenaw Peninsula small stocks of rhyolite intrude the basaltic lavas. In the Porcupine Mountains, near the Michigan-Wisconsin boundary, large masses of quartz porphyry, felsite, and other acidic rocks intrude the Keweenawan lavas. These acidic intrusions may largely explain the source of the abundant felsite, quartz porphyry, and augite syenite pebbles in the Copper Harbor conglomerate. Furthermore, recent unpublished work by a group at the Massachusetts Institute of Technology indicates that the nepheline and augite syenites near Coldwell on the northeast shore of Lake Superior are about 1,000 to 1,100 million years old. These ages are based on three determinations, a strontium-rubidium age on biotite, potassium-argon on biotite, and strontium-rubidium on potassium feldspar. These ages agree well with the 1,120 million years for the Duluth gabbro, and the syenitic intrusions may belong to the late middle Keweenawan period of intrusion of basic and acidic rocks, as indicated by geology and palaeomagnetism.

The Freda and Nonesuch pole (Fig. 29, No. 12) is significantly different from the Portage Lake and Copper Harbor poles. This marked difference would indicate that the contact between the Nonesuch and the underlying Copper Harbor represents a definite break in sedimentation, although it is possible that it represents instead a period of rapid polar wandering. Geologically the contact appears conformable, and it has always been assumed that there was no sedimentary break, only a change in facies. Consequently, in the geological nomenclature, the Portage Lake lavas are called middle Keweenawan, and the Copper Harbor and all later sediments are considered to be upper Keweenawan. However, the palaeomagnetic results indicate that the Copper Harbor is more closely related to the lavas and intrusions of the middle Keweenawan than the later sediments and should be classified as such. It is worth noting also that the Copper Harbor contains lava flows, whereas the later sediments contain no such lavas nor are they intruded by dykes and sills.

The Mamainse Point Keweenawan lavas (Fig. 29, No. 7) have long been considered to be equivalent to the Portage Lake lavas (*see* Fig. 29, No. 6). This correlation was based on petrological and structural similarities between the two formations, but there has never been any conclusive geological evidence. The palaeomagnetic data from Mamainse Point gave a pole position that is not significantly different from that of the Portage Lake pole. The circle of confidence is large enough so that the Mamainse lavas might be referred to the Duluth gabbro and Copper Harbor conglomerate, but as the pole position is closer to the Portage Lake pole and the geological evidence supports such a correlation, these Mamainse lavas may be correlated with considerable confidence with the Portage Lake lavas.

Overlying the so-called Keweenawan lavas at Alona Bay are gently dipping beds of fine-grained sandstone. Hamblin (1958), because of their similar lithology, considered these sandstones to be equivalent to the Freda sandstone. Furthermore, he suggested that they underlie the Jacobsville sandstone found in the Sault Ste. Marie area. All these conclusions are based on very scanty evidence, but they do show that the Alona Bay sandstones are different from the Jacobsville. Their palaeomagnetism supports such a conclusion and indicates a correlation with the Copper Harbor-Duluth gabbro group, because their pole position is not significantly different. The palaeomagnetic evidence is, however, slender, being based on only four samples, and more work is necessary to confirm this suggestion.

The Logan diabase dykes referred to as group II give a pole position very close to and not significantly different from the Portage Lake pole position. This can be interpreted as good evidence for correlating these dykes with the lavas. Indeed the dykes may have been the feeders for the lavas that outcrop to the southeast and northeast, on Isle Royale and the Ontario shore of Lake Superior. These dykes have long been considered to be of the same age as the Logan sills, with which they are intimately associated and to which they have a strong petrological similarity. However, some of the dykes that were studied palaeomagnetically intrude the sills and, therefore, are definitely younger. There is, however, no geological evidence to determine how much younger they may be. The palaeomagnetic evidence also indicates that the dykes are considerably younger than the sills and suggests that their petrological similarity is due not to contemporaneity but to their occurrence in the same petrological province.

## Upper Keweenawan

The palaeomagnetic results for the Freda sandstone and Nonesuch shales (Fig. 29, No. 12) and for the Eileen sandstone (Fig. 29, No. 13) give pole positions that are significantly different from one another. This is more because their circles of confidence are small, as the result of extensive sampling and a low dispersion of the remanent magnetic vectors, than that the poles are actually far apart (4 degrees in latitude and 19.5 degrees in longitude). It would appear that the pole moved about 20 degrees westwards between the end of Freda time and the beginning of Eileen time.

The results from the Orienta are rather unsatisfactory in that too many of the samples had to be discarded as obviously unstable. The remaining samples were too scattered and too few for an accurate pole position to be calculated. Despite these drawbacks the Orienta pole is not significantly different from those of the Freda and Nonesuch and not very different from that of the Eileen sandstone.

The Jacobsville sandstone near Sault Ste. Marie gives a pole position that is not significantly different from those of the Freda, Eileen, and Orienta. However, it is significantly different from the Jacobsville on Keweenaw Peninsula. This discrepancy might be because the last rocks, which occur at the top of the thickest part of the clastic wedge of Jacobsville sandstone described by Hamblin, may be younger than those at Sault Ste. Marie. A second explanation could be that there was considerable random polar wandering during Jacobsville time and that the sampling at the Sault and Keweenaw Peninsula did not cover a sufficient length of time to smooth out these random fluctuations. In general, both of the Jacobsville pole positions are sufficiently similar to the other upper Keweenawan pole positions to suggest that the Jacobsville belong to the upper Keweenawan rather than to the Cambrian, as some geologists have suggested.

The exact relations between the Jacobsville and the other upper Keweenawan sediments are unknown. Hamblin (1958) found that the Jacobsville in places contains pebbles that are similar to the Freda sandstone. He also points out that near Sault Ste. Marie there appears to be an angular unconformity between the Jacobsville and certain rocks that are like the Freda but which may, on the palaeomagnetic evidence, be correlated with the middle Keweenawan. Lithologically the Jacobsville closely resembles the Bayfield group, particularly the Orienta sandstone, and it is worth noting that neither of the two Jacobsville poles is significantly different from that of the Orienta. A consistent but by no means certain correlation would be to relate the Jacobsville to the Orienta.

The Chequamegon pole position is very different from that of any of the other upper Keweenawan poles and implies that a major break took place between the deposition of the Orienta and that of the Chequamegon. Such a break might be either a sudden shift in pole position or a hiatus of non-sedimentation.

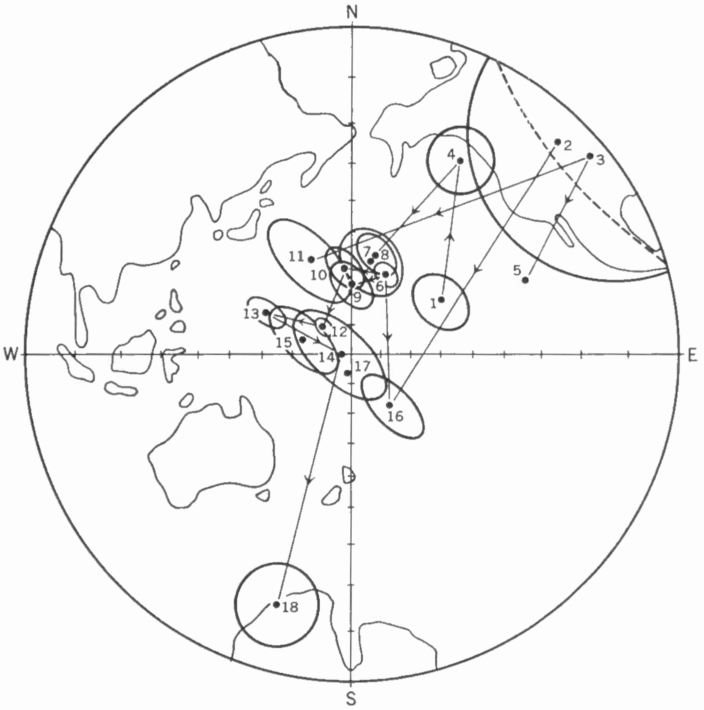

*Figure 29.* *The Keweenawan pole positions. Numbers refer to the formations listed in Table XVIII. Circle of confidence for sandstone dykes (5) not drawn; it is so large that diagram would be confused.*

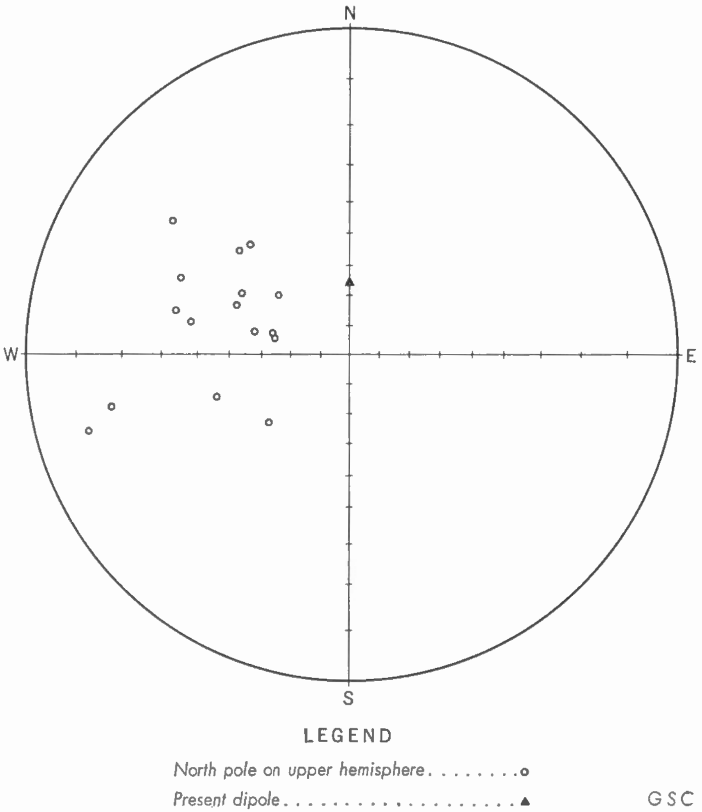

*Figure 30.* *Directions of magnetization of rocks from Grenville province corrected to 45.0°N and 78.5°W.*

## Table XIX

### *Correlation of Keweenawan Units*

| | | | |
|---|---|---|---|
| Chequamegon | | | |
| Orienta | Jacobsville on Keweenaw Peninsula | | Jacobsville at Sault Ste. Marie |
| Eileen sandstone | | | |
| Freda and Nonesuch | | | Freda (?) at Sault Ste. Marie |
| Duluth gabbro = 1,120 my<br>Copper Harbor | | Osler lavas<br>Diabase dykes | |
| Portage Lake lavas | | | Mamainse Point lavas |
| Baraga county diabase dykes 1,400 my | | Logan sills | Alona Bay lavas |
| | | Sibley series | |
| Animikie 1,700 my age 2,100 my | | Animikie | |
| Northern Michigan, Wisconsin and Minnesota | | Western Ontario | Sault Ste. Marie area |

## *Chapter VI*

# CORRELATION WITH OTHER LATE PRECAMBRIAN FORMATIONS OF NORTH AMERICA

## Summary of Results

Recently more palaeomagnetic data for the late Precambrian have become available for the North American continent. In this section these data are discussed and related to those for the Keweenawan.

Collinson and Runcorn (1959) have recently obtained palaeomagnetic data from the Belt series of northwestern United States. Their results from the various formations of the Belt are tabulated in Table XX, and the mean pole position has been listed in Table XXII and plotted in Figure 31.

**Table XX**

*Palaeomagnetic Results from Belt Series*

| Formation | D | I | | N | Pole position | Collecting site |
|---|---|---|---|---|---|---|
| Macnamara | 26 | −43 | 4 | 20 | 15S, 138W | 47N, 114W |
| Miller Peak | 234 | +30 | 7 | 14 | 11S, 168W | 47N, 114W |
| Spokane | 220 | +52 | 4 | 39 | 2S, 147W | 47N, 114W |
| Grinnell | 225 | +48 | 6 | 16 | 4S, 153W | 47N, 114W |
| Appekunny | 223 | +29 | 6 | 15 | 17S, 158W | 47N, 114W |
| Mean direction | 227 | +41 | — | — | 8S, 157W | — — |

Hood (1958) obtained data for the Sudbury norite and certain gabbro bodies in the Bancroft region of central Ontario. He found that the palaeomagnetic results for the North and South Ranges of the Sudbury norite lopolith give directions of magnetization that differ by 40 degrees. He could explain this discrepancy by assuming that the lopolith was folded so that its margins became more steeply inclined towards its centre. He computed a pole position of 38.4°N and 99.4°W by combining and by assuming that each margin had been tilted 20 degrees since cooling. However, the dips are much steeper on the South Range, where in some places the beds have even been overturned and where there is evidence of considerable faulting, whereas at Levack, where he collected his samples from the North Range, the base of the norite dips at only 35°S. It, therefore, seems likely that the Levack samples have been structurally tilted less than those from the South Range, and possibly they have not been tilted at all. The pole position computed by Hood for the Levack samples, assuming that they have not been tilted, is 64.0°N and 140.7°W. This pole position is listed in Table XXII and plotted in Figure 31.

Hood also measured two groups of gabbro bodies from the Bancroft area, the Boulter gabbro and the Umfraville-Thanet gabbros. His results for these rock units are also listed and plotted in the same table and figure. It should be noted that the pole plotted for the Umfraville-Thanet gabbro is in the southern Pacific whereas the corresponding pole listed by Hood is in the northern Atlantic. This discrepancy is only apparent as the pole listed in this paper is the anti-pole to that of Hood's.

Martinez, Howell, and Statham (1958) have reported on the palaeomagnetism of the flat-lying late Precambrian Hazel formation of Texas, for which they calculated a pole position of 40.0°N and 175.0°W.

Recently Collinson and Runcorn (1959) have obtained results for the Hakatai shale from the Grand Canyon of Arizona. Their computed pole position is 27.0°N and 175.0°E, which differs somewhat from earlier results for this formation. Only future work can decide whether this discrepancy is due to age difference, isothermal remanence, or some other secondary magnetization. In the same paper

## Table XXI

*Results for Rocks of Grenville Province Corrected to 45° N and 78.5° W of Greenwich*

| Specimen | D | I | Description of locality |
|---|---|---|---|
| M 12 | 283.5 | −32.5 | Collected by R. Mitra in 1954 from the centre of Digby tp., Victoria co., Ont. at lat. 45°47′ N and long. 78°55′ W; samples were taken in area between Fishog Lake and North Fishog Lake, over part of this area is a prominent negative aeromagnetic anomaly. |
| M 13 | 299.5 | −49.0 | |
| M 23 | 313.0 | −41.0 | |
| M 24 | 293.5 | −49.5 | |
| M 29 | 318.5 | −41.5 | |
| M 30 | 257.5 | −16.5 | |
| M 40 | 253.5 | −10.5 | |
| M 41 | 306.5 | −22.0 | |
| M 42 | 308.5 | −59.0 | |
| M 43 | 281.0 | −64.0 | |
| BAN. 1 | 283.0 | −57.0 | Collected about a mile south of Allen Lake in Harcourt tp., Haliburton co. at lat. 45°5′ N and long. 78° 15′ W; locality at centre of rather intense negative aeromagnetic anomaly. |
| BAN. 2 | 282.5 | −63.5 | |
| W 1 | 281.0 | −37.5 | Collected from an intense negative aeromagnetic anomaly about 2¼ miles east of Alcove in Wakefield tp., Gatineau co., at lat. 45°41′ N and long. 75°53′ W. |
| W 2 | 230.0 | −54.0 | |
| W 3 | 294.0 | −30.5 | |
| W 4 | 252.5 | −44.0 | |

## Table XXII

### *Pole Positions During Late Precambrian*

| [1]Formation | Pole position |
|---|---|
| 25. Chequamegon | 68.0S, 131.5E |
| 24. Umfraville and Thanet | 34.2S, 157.8E |
| 23. Lodore | 27.0S, 169.0E |
| 22. Jacobsville (Keweenaw) | 17.5S, 166.0W |
| 21. Grenville | 7.0S, 162.0E |
| 20. Mean Jacobsville | 6.5S, 179.0E |
| 19. Orienta I | 0.0 , 176.0E |
| 18. Jacobsville (Sault) | 5.0N, 164.0E |
| 17. Eileen | 13.0N, 150.5E |
| 16. Freda and Nonesuch | 9.0N, 170.0E |
| 15. Hakatai | 27.0N, 175.0E |
| 14. Sault Freda (?) | 31.0N, 164.5E |
| 13. Copper Harbor | 29.0N, 176.5E |
| 12. Duluth gabbro | 23.5N, 179.5W |
| 11. Mamainse Point lavas | 31.5N, 173.5W |
| 10. Portage Lake | 26.5N, 168.5W |
| 9. Logan diabase II | 33.0N, 171.5W |
| 8. Hazel | 49.0N, 175.0W |
| 7. Boulter | 42.6N, 156.8W |
| 6. Levack | 64.0N, 140.7W |
| 5. Logan diabase I | 53.5N, 130.0W |
| 4. Baraga co. dykes | 45.0N, 99.5W |
| 3. Alona Bay lavas | 39.0N, 95.0W |
| 2. Sibley series | 16.0N, 149.0W |
| 1. Belt series | 8.0S, 157.0W |

[1] Numbers refer to pole positions plotted on Figure 31.

Collinson and Runcorn reported the results for the Lodore formation, which is considered by some geologists to be Cambrian. However, its age cannot be determined exactly by geological methods. The Lodore is overlain, apparently conformably, by Cambrian strata and underlain by non-fossiliferous red beds that are considered to be Precambrian. The Lodore itself is unfossiliferous and has apparently been correlated with the Cambrian only on its lithological similarities to certain distant Cambrian rocks and its apparent conformity with known Cambrian rocks. Its age, therefore, is very much in doubt. Palaeomagnetism would indicate that it may be Precambrian.

Finally, the author can report on certain very recent results from the Grenville province. These rocks were collected from three different localities, well separated from one another, described briefly in Table XXI. Figure 30 is a stereographic plot of the results, whose mean direction is 286.5 degrees east of north and —44.5 degrees, which gives a pole position at 162.0°E and 7.0°S. The rocks from Digby township are granite-gneisses, those from Harcourt township are pyroxenites, and those from Wakefield are basic syenitic gneisses. The Digby township samples when first measured gave somewhat scattered and inconsistent results. However, after A. C. demagnetization in a peak field of about 200 gauss there was a great improvement in the consistency of magnetization. Five samples, M9, M10, M33, M34 and M35, were not included in the results given in Table XXI and Figure 30, because they showed internal inconsistency and were very weak.

*Figure 31.* **Late Precambrian pole positions.**

## Discussion of Results

On Figure 31 the pole positions calculated from these additional Late Precambrian results have been plotted (crosses) together with the Keweenawan results (full circles) already described. The sandstone-dykes pole has too large a circle of confidence to be useful and is omitted. A generalized polar wandering curve has been drawn on the basis of the Keweenawan results. It can be seen that the additional results fall closely along this generalized path. Examination of Figure 31 suggests the following age relationships. The pole position for the Belt series is not far from that of the lower Keweenawan Sibley series (No. 2) or those of the upper Keweenawan formations (Nos. 16-20). The palaeomagnetic evidence therefore is not sufficiently critical to allow positive correlation. Reasons for correlating the Belt with the Sibley are given later. The Levack samples from the Sudbury norite (No. 6) may be correlated approximately with the Logan sills (No. 5). The Boulter gabbro (pole No. 7) and Hazel formation (pole No. 8) may be roughly the same age and may have been formed after the intrusion of the Sudbury norite and Logan sills but before the extrusion of the Portage Lake lavas (No. 10). The Hakatai shale (No. 15) should be correlated with the middle Keweenawan lavas and intrusions. The Grenville gneisses (No. 21) of Digby, Harcourt, and Wakefield townships may have been metamorphosed at about the time the Nonesuch, Freda, and Orienta sediments were deposited. The Lodore formation (No. 23) and the Umfraville-Thanet gabbro (No. 24) may have been formed at about the same time and after the deposition of the Nonesuch, Freda, Eileen, and Orienta sedimentary rocks and before that of the Chequamegon sandstone.

The Belt series is overlain unconformably by Lower Cambrian strata, but the length of time of erosion or non-deposition represented by this unconformity has been the subject of considerable discussion. In many places there is little angular unconformity between the Belt and the overlying Cambrian, and the Belt itself contains evidence of primitive life. For these reasons geologists have stated that the Belt was very close in age to the Cambrian if not actually part of it. However, more detailed geological work has shown that the unconformity is a major one. The Cambrian transgresses across practically all the formations of the Belt series, and in southeastern British Columbia the equivalent of the Belt is overlain unconformably by thousands of feet of the Windermere system, which is in turn overlain apparently conformably by Lower Cambrian rocks (Reesor, 1957). Further study may show that this apparent conformity is actually an unconformity. The basal member of the Windermere system is the Toby conglomerate, a fairly extensive formation of very variable thickness. Reesor (op. cit.), with respect to the unconformity at the top of the Belt series stated: "It is in any case evident from the unconformity represented by the presence of the Toby conglomerate at the base of the Windermere system that the Purcell basin was deformed at the end of Purcell (Belt) time. No evidence exists that this deformation was intense, but it was rather open folding and upwarping." Further on Reesor discussed the amount of erosion that took place as a consequence of this deformation and concluded: "The present distribution of the Cranbrook indicates that many thousands of feet of sedimentary rocks (all of Upper Purcell and perhaps one-half of Lower Purcell) were eroded from the ancestral Purcell Mountains between the end of Purcell (Belt) time and before the beginning of the Cambrian." The thickness of the sediments removed has been estimated to be about 20,000 feet. The purely geological evidence indicates therefore that a considerable time interval must have elapsed between the end of the Belt and the beginning of the Cambrian.

The geological evidence has recently been supplemented by several age determinations. Eckelmann and Miller (1956) dated at 1,190 million years a pitchblende vein cutting the St. Regis quartzite, as a member of the Ravalli group of lower Belt age, at the Sunshine Mine near Coeur d'Alene, Idaho. Farquhar and Cumming (1954) obtained by the method of lead isotope ratios an age of 1,030 ± 290 million years for the Sullivan Kimberley ore deposit, the host rock of which belongs to the Purcell system. Goldich, *et al.* (1959) have recently measured the age of an illite-bearing shale in the Siyeh formation of the Belt series at Cogan Pass, Glacier Park, Montana. They obtained a potassium-argon age of 740 million years and a strontium-rubidium age of 780 million years for the formation of the illite. These two ages agree with each other and are probably good indicators of the time when the illite was last crystallized. However, the ages are considerably less than those determined by Farquhar and Cumming (1954), and Eckelmann and Miller (1956) and may represent the time of some metamorphism that occurred considerably after the deposition of the Belt sediments. The minimum age of the Belt is, therefore, probably about 1,200 million years, and the true age could be considerably greater. Because of its estimated age it is probable that the Belt series is older than middle Keweenawan Duluth gabbro, which has been dated as 1,120 million years old, and may well be correlated with the lower Keweenawan Sibley series, the alternative correlation suggested by the palaeomagnetic evidence.

Aldrich and Wetherill (1958) have published ages for gneisses from Zoroaster, Grand Canyon, Arizona. They obtained a potassium-argon age of 1,350 million years and a strontium-rubidium age of 1,370 million years. As these gneisses are overlain with a profound unconformity by the Grand Canyon series, which includes the Hakatai shale as one of its formations, the Hakatai shale must be less than 1,350 million years old. Such an age would be consistent with the palaeomagnetic evidence.

On the basis of ages determined by the lead isotope ratios in galenas it has been suggested (Cumming, *et al.,* 1955) that the Sudbury norite is 1,200 million years or older. This age would be in general agreement with the palaeomagnetic inferences, which suggest that the norite is older than the Duluth gabbro, which has already been shown to be about 1,120 million years old. However, the lead isotope age for the norite is open to considerable doubt, because many of the isotopic ratios in the galenas are anomalous.

The Grenville metamorphism has been fairly well dated as having occurred about 1,050 million years ago. If the Grenville samples cited in this paper acquired their magnetization at the time of this metamorphism, the age determination agrees with the palaeomagnetic evidence, which indicates that the Grenville samples were magnetized after the Duluth gabbro. It is interesting to note that palaeomagnetism indicates that most of the upper Keweenawan sediments were deposited more or less contemporaneously with the Grenville metamorphism and orogeny. This suggestion is in accordance with Wilson's (Kuiper, 1954) hypothesis that the deposition of the Keweenawan was the result of Grenville mountain-building and represents a post-orogenic secondary sedimentation.

## Conclusion

In this paper palaeomagnetic data concerning the late Precambrian Keweenawan of the Lake Superior region that has accumulated during the past 4 years, have been presented, with brief descriptions of the techniques of measurement and 'magnetic washing'. A detailed account of the palaeomagnetic results have been given in order to facilitate the evaluation of the method as a means of correlation and with the hope that such an account may provide a starting-point from which to continue and extend this work. In general the method appears to offer considerable promise as a geological tool. The palaeomagnetic deductions are in complete accord with the known geological facts and the isotopic age determinations. The only apparent contradictions between palaeomagnetism and geology occur where geological deductions are based on insufficient and indeterminate evidence. Palaeomagnetism may become of comparable importance to isotopic age determination, and the two together, combined with standard geological procedures, may enable problems that today are intractable to be unravelled.

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