DESIGN AND CALIBRATION OF AN ASTATIC MAGNETOMETER AND THE STUDY OF REMANENT MAGNETIZATION OF SOME NEWFOUNDLAND ROCKS

GUMMULURU S. MURTHY

DESIGN AND CALIBRATION OF AN ASTATIC MAGNETOMETER
AND THE STUDY OF REMANENT MAGNETIZATION
OF SOME NEWFOUNDLAND ROCKS

BY

© GUMMULURU S. MURTHY, M.Sc

"Submitted in partial fulfilment of the
requirements for the degree of Master of Science,
Memorial University of Newfoundland,
September 12, 1966".

ABSTRACT

An astatic magnetometer utilizing transversely magnetized magnets of large specific dipole moment has been designed and calibrated. With a period T = 4.6 seconds, P/I = 1.35×10^3 c.g.s. units, and an astaticism of 6×10^3 or more, the performance of the instrument compares well with that of more sensitive astatic magnetometers described in the literature. The present reciprocal sensitivity of 3.9×10^{-7} oe/mm scale deflection is sufficient for the accurate measurement of remanent magnetization of specimens with minimum intensity $J = 1 \times 10^{-5}$ e.m.u./cm. A tenfold increase in sensitivity, permitting measurement with relatively short period ($T < 15$ sec.) of the more weakly magnetic sedimentary rocks, is easily accomplished by changing the suspension, but has not been carried out pending removal of the instrument to a magnetically less disturbed site. At the present location on the second floor of a university science building, serious short-period mechanical vibrations also presented a problem at first, but these have been virtually eliminated by means of a specially designed, top-heavy spring suspension for the magnetometer.

The magnetometer has been used to measure the natural remanent magnetization of rocks collected in 1965 in three areas of Newfoundland. A detailed study of 25 samples of flat-lying

basalt from two sites on the Labrador Coast yielded a mean direction of NRM with azimuth, A = 21.2°; dip, D = 70.5°; and radius of circle of confidence, $\alpha_{95} = 7.0°$. This direction is too close to that of the present geomagnetic field or axial dipole field at these sites to exclude the possible presence of a significant unstable component of magnetization. Initial stability tests using only alternating field demagnetization are still inconclusive, and formulation of palaeomagnetic conclusions from the Labrador results, which could have a bearing on such problems as the age of the basalts or the postulated rotation of Newfoundland, must await further tests, including thermal demagnetization.

Some preliminary palaeomagnetic measurements were carried out on sedimentary rock samples collected off Notre Dame Bay on the island of Newfoundland, and on the Grand Banks of Newfoundland. The former collection consisted of grey and red argillites of Ordovician age from four sites, while the latter were three samples of microbreccia; these had been oriented under water and recovered by divers from the Virgin Rocks shoals, which are 170 miles east of Cape Race, on the Avalon Peninsula and probably of late Precambrian age. The results, which are still tentative, are discussed in the light of earlier palaeomagnetic work in Newfoundland.

CONTENTS

INTRODUCTION

Palaeomagnetism is fast developing into one of the most widely studied branches of geophysics. The interest in this subject originates mainly from the fact that knowledge of palaeomagnetic directions of rocks can be effectively used to solve many physical, geological and minerological problems.

One of the properties usually exhibited by a rock is its behaviour as a magnet with a North and a South pole and a magnetic axis. The magnetization in a rock can usually be expressed in terms of two components, the induced magnetization and the remanent magnetization. The former requires the presence of an external field while the latter can be present in the absence of such a field. It is this remanent or permanent magnetization which is of importance in palaeomagnetism.

Most rocks owe their magnetism to the presence of one or more oxides of iron, while in some cases iron sulphides are also of importance. Usually these minerals constitute only a small proportion of the rock volume, being distributed in a groundmass of paramagnetic or diamagnetic minerals. Magnetite (Fe_3O_4), titanomagnetite (solid solution between Fe_3O_4 and $TiFe_2O_4$), maghemite (γ-Fe_2O_3), hematite (α-Fe_2O_3), and ilmenite ($TiFeO_3$) are the oxides of iron most important in rock magnetism, while pyrrhotite (FeS_{1+x}) is the only important sulphide. In all cases the significance of these minerals, or of the solid solutions between some of the oxides, lies in their property of ferromagnetism which enables them to have remanent magnetism in the absence of an external field.

An important property of ferromagnetic materials is that they occur in the form of crystals. The ferromagnetic character of the material is then derived from strong interaction of electron spins among neighbouring ions in the crystal lattice, which tends to align all the spin magnetic moment vectors against the randomizing effect of thermal agitation. Then, according to the quantum mechanical interpretation, the exchange energy will be a minimum when the spins of the neighbouring ions or atoms become parallel or antiparallel. Thus there will be a net magnetization (the spontaneous magnetization) in the absence of an external field. When the temperature is raised the spins are increasingly deflected from their close alignment; hence the spontaneous magnetization decreases, falling to zero at the Curie temperature, T_c; this is a characteristic parameter of ferromagnetic materials, which become paramagnetic at temperatures above T_c.

Despite the property of spontaneous magnetization, ferromagnetic materials nearly always need an external field to acquire a remanent magnetization. This phenomenon can be explained by the Domain Theory, according to which even a single ferromagnetic crystal usually consists of numerous magnetic domains separated by discontinuous boundaries. Thus, while individual domains have parallel spin alignment, the vector sum of the moments of a large number of domains is zero. An external field gives the material a net magnetization by deflecting the average spin moment towards the field direction.

With or without an external field, however, the arrangement of magnetic moments in the crystal depends upon the condition that the total energy per unit volume must be a minimum. If the entire volume of the crystal or grain were a single domain, the <u>magnetostatic energy</u> would become significant because of the dominant effect of free poles. This tendency is opposed by the formation of domains having the size and shape consistent with minimum total energy. Due to the <u>magneto-crystalline energy</u>, however, the parallel spin moments in each domain tend to lie along the crystal axis of easiest magnetization. This will further modify the size and shape of the domains, tending to enlarge those whose magnetization lies in the direction of the external field, and to reduce the size of the others. Finally, a considerable amount of <u>magnetoelastic energy</u>, arising from internal stress, may be stored in the material, so that in the end four energy forms, exchange, magnetostatic, magnetocrystalline and magnetoelastic energy contribute to determine the optimum number, size and shape of domains in a ferromagnetic substance.

Actually, the magnetic properties of the minerals important in rock magnetism are largely derived from ferri-magnetism, a phenomenon closely related to ferromagnetism. Here the magnetic ions occupy two sublattices, A and B, in the unit cell of the crystal, where the spins in A are directed in opposition to those in B because of strong negative interaction between the two spin systems. The result is an antiparallel spin arrangement, but with a net magnetic moment in one or the other direction, since the number of ions as well as their individual moments are different in A and B. Magnetite, titano-magnetite, maghemite, pyrrhotite, and ilmenite-hematite solid solutions in a certain compositional range, all exhibit ferri-magnetism.

If the energy supplied by the applied field is not greater than the magnetic energy barriers, the direction of magnetization will return to its original position when the field is removed. Thus the reversible or induced magnetization will result.

However, a characteristic property of all ferromagnetic substances is magnetic hysteresis: their intensity of magnet-ization, J, reaches a maximum value, J_s, at saturation in a finite magnetizing field, but does not disappear when the field is again reduced to zero. Instead, the domain vectors retain a preferential alignment that corresponds to the remanent magnet-ization, J_r, whose magnitude may be a substantial proportion of J_s. When the field is increased again, but in opposition to its previous direction, J decreases further until it vanishes for a field $-H_c$, the coercive force.

Rocks may acquire their remanent magnetization in one of several ways depending on the method of formation of the rock and its subsequent history. Igneous rocks, which are formed by the cooling of liquid magma, tend to acquire most of their magnetization as they cool through the Curie temperatures of their ferromagnetic constituents. Laboratory experiments by numerous workers (Irving 1964) indicate that the direction of magnetization acquired by this process, known as thermoremanent magnetization (TRM) in most cases lies accurately in the direction of the applied field. However, in rocks having shape or crystal anisotropy, the TRM may not be along the ambient field, in which case they make palaeomagnetic interpretations more difficult. Anisotropy tends to be negligible in igneous rocks, which have no grain structure and prominently contain the isotropic crystals of magnetite, titanomagnetite, or maghemite. It may not be negligible in rocks whose main ferromagnetic component has strong crystalline anisotropy (e.g. the ilmenite-hematite system, or pyrrhotite), or in certain sedimentary or metamorphic rocks where shape anisotropy may result in a preferential alignment of grains related to rock structure.

Néel (1955) proposed a theory of the TRM for single-domain particles. In Néel's theory, the magnetic moment of the single grain can lie along one of two mutually opposite directions consistent with minimun energy per unit volume of the grain. Separating these directions is an energy barrier given by

$$E = \frac{vH_cJ_s}{2}$$

where v is the volume of the grain, H_c is the coercive force and J_s is the spontaneous magnetization of the mineral. If E exceeds the thermal energy kT (k = Boltzmann's constant, T = Absolute temperature), no change of the magnetic direction across the barrier will result. When the temperature is sufficiently high, or the volume sufficiently small, however, the magnetic moments can cross the energy barrier, causing all magnetic vectors to have a single orientation. The combined effect due to all the oriented domains results in a value J_o of the initial magnetization.

After a given time t, this value will decay to J_R where

$$J_R = J_o e^{-t/\tau_o}$$

and τ_o is called the relaxation time. Néel (1955) defined τ_o as

$$\frac{1}{\tau_o} = A\left(\frac{v}{T}\right)^{1/2} e^{-vH_cJ_s/2kT}$$

$$\frac{1}{\tau_o} = A\left(\frac{v}{T}\right)^{1/2} e^{-\gamma v/T}$$

where constants A and γ depend on the elastic and magnetic properties of the mineral. As these constants are known for iron, Néel showed that τ_o , v/T and grain diameters at room temperature are related as follows in the case of iron:

v/T	τ_o	Grain diameter (at room temperature)
3.2×10^{-21}	0.1 secs	120 Å
7.0×10^{-21}	340 years	160 Å
9.6×10^{-21}	3.4×10^{9} years	180 Å

Thus the direction of magnetization in a grain with diameter less than 120 Å and at room temperature is quickly changed by the thermal fluctuations, and the application of a weak field to a number of such grains causes a net magnetization in the direction of the field. On the other hand, an increase in v/T by a factor of 3 increases τ_o by a factor of 10^{17}. Stacey (1963) proposed a theory of TRM for multi-domain grains.

One way in which sedimentary rocks may acquire their remanent magnetization is by the preferential alignment of the ferromagnetic mineral grains (which along with non-magnetic particles, have been derived from older rocks by erosion) along the ambient field during their deposition, say in water. This is called detrital (or depositional) remanent magnetization (DRM). However, Griffiths, King and Wright (1957) point out that the magnetic inclination may deviate systematically from that of the ambient field due to rolling of particles over others at the sediment surface as this process is affected by shape anisotropy.

There are also other types of remanent magnetism which a rock can acquire, such as <u>chemical remanent magnetization</u> (CRM), <u>isothermal remanent magnetization</u> (IRM) and <u>viscous remanent magnetization</u> (VRM). Cox and Doell (1960), Nagata (1961), and Irving (1964) have discussed in detail these various forms of natural remanent magnetization (NRM) of rocks.

The earlier classic work in palaeomagnetism by Chevallier (1925) indicated that the remanent magnetism of lava flows on Mt. Etna was parallel to the earth's magnetic field as measured at nearby observatories at the time the flows erupted. Thus Chevallier established clearly that the rocks which he studied acquired their remanent magnetism at the time of their formation. Other pioneers, such as Koenigsberger (1930), demonstrated repeatedly in the laboratory that the remanent magnetization of rocks heated above the Curie points of their ferromagnetic constituents and cooled in an ambient field nearly always had the same direction and sense as that field. These early investigations served to establish confidence in the ultimate usefulness of the NRM of rocks as a fossil compass. This confidence has been justified by more recent studies, which have also increasingly emphasized, however, that reliable conclusions from rock magnetism can be drawn only if the "stability" of the NRM, as defined below, can be clearly established.

The stability of rock magnetism is its resistance to change under given conditions; thus the NRM of a rock is said to be stable if it, or some identifiable component of it, has not changed its direction significantly in the presence of forces tending to alter it during the rock's history. Both field and laboratory tests can be performed to study the stability. The field tests include Graham's classic fold test and the conglomerate test (Graham, 1949). In the laboratory, the stability of NRM is tested by one or more demagnetization procedures, the most widely used being alternating field demagnetization (Thellier and Rimbert, 1954) and thermal demagnetization (Thellier, 1938). In the former method, the rocks are partly demagnetized in a smoothly decreasing alternating field, with the aim of randomizing the directions of the magnetic moments in all domains having a coercive force less than the highest value of the demagnetizing field: this would leave intact the magnetization due to the "hard" (high coercive force) domains, which can often be assumed to have preserved the direction of the ambient field at the time the rock was formed.

In thermal demagnetization the rock is heated to a certain temperature T_1 and cooled in a zero field to atmospheric temperature, T_0 , the direction and intensity of remanent magnetization being measured before and after heat treatment. This process will destroy any magnetic component with blocking temperature, $T_B < T_1$, where T_B is the temperature at which thermal agitation becomes effective in randomizing the domain moments of that component in a short time compared with the time of a measurement. Treatment is then repeated by heating to a temperature $T_2 > T_1$ and again cooling to T_0 in a zero field, and so on, where T_1, $T_2 \ldots T_n < T_c$, and T_c corresponds to the component with the highest Curie temperature. The effectiveness of thermal demagnetization depends upon the fact that in many rocks the most stable (or "hardest") magnetic component is due to a TRM acquired in a weak field just below the Curie temperature, which remains after unwanted, less stable components (such as an IRM or partial TRM acquired at relatively low temperature a significantly long time after the rock was formed) have been removed by the treatment. An alternative thermal demagnetization technique consists of taking measurements during continuous heating of the sample to a high temperature, and calculating the direction and magnitude of the components destroyed in each temperature interval.

Knowledge of the NRM of rocks has many useful applications. The evidence can be interpreted in terms of the geophysical hypotheses of continental drift and polar wandering. They have also found application in solving local tectonic problems, as well as in age correlation of rocks. An important application arises from reversals in polarity, which are a frequent occurrence in rock formations of various ages: these have been cited by many workers as evidence for reversals of the geomagnetic field during geological times (Hospers, 1955; Cox, Doell and Dalrymple, 1963), but an alternate possibility is that in some cases physico-chemical changes in the rock may cause it to acquire a magnetization directed in opposition to the acting field (Neel, 1951; Uyeda, 1958).

However, the measurement of remanent magnetization is essentially difficult. In general, measurements are carried out either electromagnetically or by means of an astatic magnetometer. Electromagnetic apparatus of the "rock generator" type was pioneered by Johnson and McNish (1938) of the Carnegie Institute of Washington. The astatic magnetometer was first used for the measurement of permanent magnetization of rocks by Japanese workers (reported in Nagata 1953). Blackett (1952) considerably improved the design of the astatic magnetometer and achieved a high sensitivity.

The Physics Department of Memorial University has initiated a comprehensive project in rock magnetism, centered on studies of the natural remanence, magnetic stability and other magnetic properties of rocks from Newfoundland and adjacent areas. As part of this endeavour, the author has designed and calibrated an astatic magnetometer of fairly high sensitivity, modifying Blackett's design in some respects. During the summer of 1965, the author collected rock samples from various parts of Newfoundland and Labrador. The natural remanent magnetization of these rocks has been studied in detail and some possible interpretations have been suggested, subject to the limitation imposed by the fact that equipment for stability testing of the rocks became available only during the last stages of this work.

The subject matter of this thesis is divided as follows:

Chapter 1 deals with the design and construction of the astatic magnetometer, and Chapter 2 with its calibration and performance.

Chapter 3 deals with the method of measurement with the astatic magnetometer.

Results of measurements of the magnetization of basalts exposed in Labrador are presented and discussed in Chapter 4.

Chapter 5 deals with a discussion of preliminary magnetic measurements of some Ordovician argillites from the Springdale area of Newfoundland.

The palaeomagnetic directions of a few samples of sedimentary breccia from the Virgin rocks shoals of the Grand Banks of Newfoundland, are discussed in Chapter 6, mainly to illustrate the procedure of measuring the remanent magnetism of cubical specimens.

A summary and conclusions are presented at the end of this thesis.

CHAPTER 1

DESIGN AND CONSTRUCTION OF THE ASTATIC MAGNETOMETER

1.1 Preliminary remarks

Methods for the measurement of magnetic fields associated with igneous and sedimentary rocks have been known for a long time, but accurate measurements of the direction of natural remanent magnetization of weekly magnetized rocks were made possible only in recent times, after the design of very sensitive magnetometers. Of these, two basic instruments, the spinner magnetometer and the astatic magnetometer, are of importance.

The spinner magnetometer, a development of the "rock generator" of Johnson and McNish (1938), and advanced by Bruckshaw and Robertson (1948) and others, works on the principle that a rotating magnetic dipole placed on the axis of a coil will induce an e.m.f. that is proportional to the frequency of rotation. In practice a double coil is used, wound in such a way that a uniformly changing magnetic field will induce no e.m.f., while a spinning rock specimen on the axis of the double coil can be approximated to a rotating dipole that will induce a net e.m.f. because of its greater proximity to the inner coil. This system effectively reduces external electromagnetic disturbances to a small value, while the combination of a large spinning frequency and high amplification of the induced signal enables one to detect even the feeble magnetic moments observed in sedimentary rocks. The direction of this remanence is then found by comparing the phase of the e.m.f. it sets up in the coil system with that of a suitable reference e.m.f.

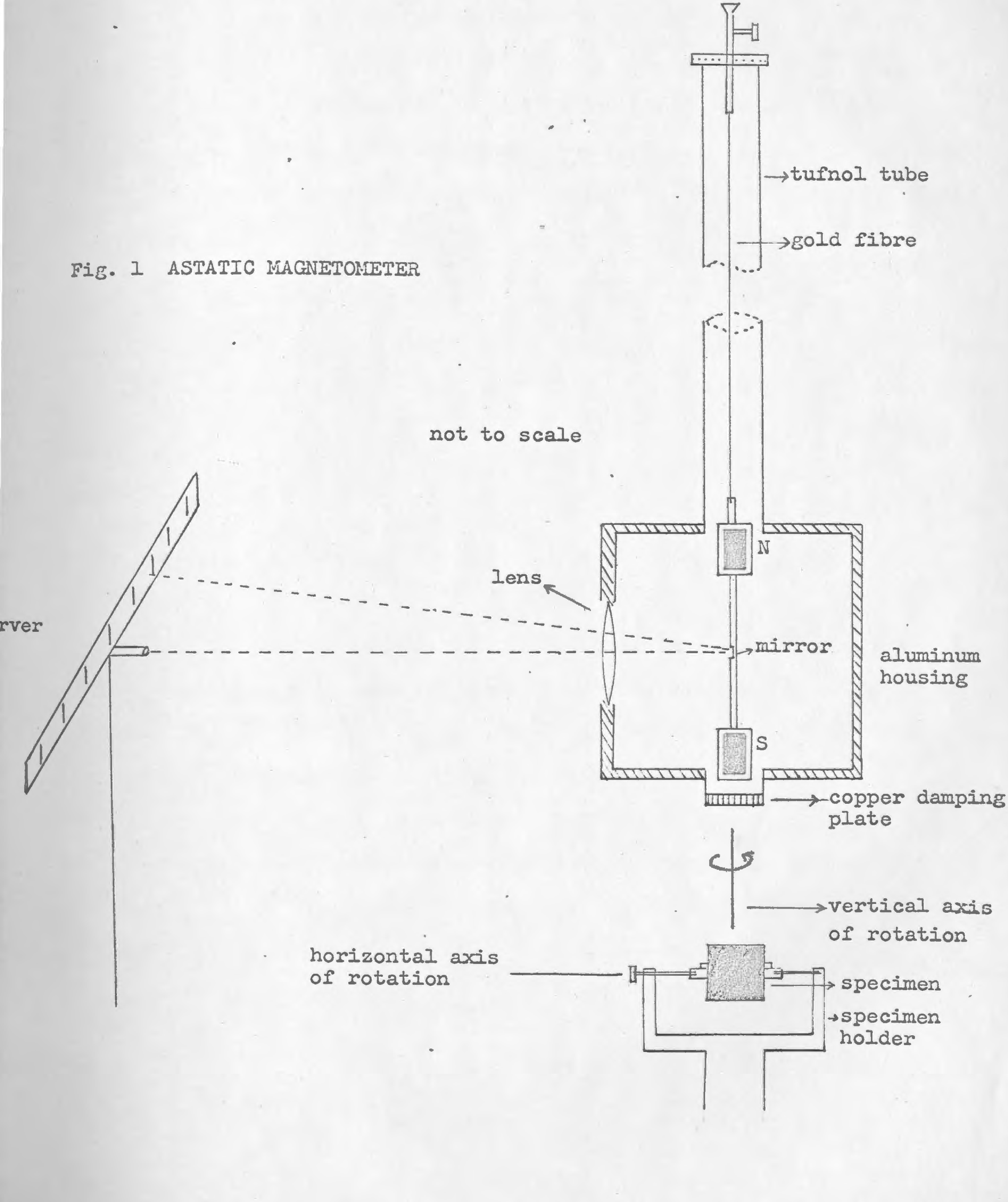
The other commonly used instrument is the astatic magnetometer, of which the pioneering application to rock magnetism is due to Blackett (1952). The author has set up an astatic magnetometer of similar general type to that of Blackett at the Physics Department of the Memorial University. The theoretical and experimental design of the above instrument forms the subject matter of this chapter.

1.2 Principle

An astatic magnetometer (Fig. 1) essentially consists of two small magnets fixed in a light vertical frame which is suspended by a thin fibre. The two magnets, of nearly equal magnetic moments, are fixed in the frame one above the other such that their directions of magnetization are horizontal and antiparallel. A small mirror, usually attached near the centre of the frame, reflects an optical beam to a scale conveniently placed at one to several meters from the system, to permit measurements of the angular deflections of the astatic system about the vertical axis to be carried out with a large magnification.

A perfect astatic system, i.e. the one in which the two magnets have exactly equal and opposite moments, should not

Fig. 1 ASTATIC MAGNETOMETER



be deflected by a changing uniform magnetic field. In a non-uniform magnetic field, the deflection of the system is proportional to the difference between the mean horizontal fields acting on the top and bottom magnets, respectively, and is thus a measure of the vertical gradient dH/dZ of the horizontal field.

Many previous research workers, Blackett (1952), Collinson and Creer (1960), Nagata (1961) and Roy (1963), have discussed the design of astatic magnetometers of high sensitivity. Even though the basic design of the magnetometer described here resembles that of the other magnetometers, there is still room for improvement regarding various details of the design, and these will now be discussed with reference to the present instrument.

1.3 Theoretical design of the astatic magnetometer

The total moment of inertia, I, of the astatic system is

$$I = 2I_o + I_1 \qquad (1.1)$$

where I_o and I_1 are the moments of inertia of a single magnet, and of the frame including the mirror, respectively. This may also be written as

$$I = \alpha I_o \qquad (1.2)$$

where α is the ratio of the moment of inertia of the system to that of the individual magnet. The undamped period of the suspended system is

$$T = 2\pi\sqrt{\frac{I}{\sigma}} \qquad (1.3)$$

where σ is the torsion constant of the suspension fibre. Since the system is operated at near critical damping, the period T of the system may be taken approximately as the time required for a single observation. The deflection, θ, (in radians) of such a system due to a magnetic field H acting at right angles to the lower magnet of dipole moment P, is

$$\theta = HP/\sigma \qquad (1.4)$$

From the above considerations, the sensitivity of the system is given by

$$\theta/H = T^2/4\pi^2 \cdot \frac{P}{I} \qquad (1.5)$$

The sensitivity is thus proportional to the square of the period for a given magnet system and to P/I for a given period.

Thus there are two possible ways of attaining a high sensitivity: (1) by choosing a long period; or (2) by designing the magnet system for the maximum possible value of P/I. However, certain practical considerations restrict the period to values below some maximum. First, at critical damping, the period, and therefore the time taken for a single measurement, would become excessive, if it exceeds, say, 30 seconds. Secondly, the variation ΔH of the local value of the geomagnetic field in the time of observation T, may be written as

$$\Delta H = T\frac{dH}{dt} + \frac{1}{2}T^2\frac{d^2H}{dt^2} + \cdots \qquad (1.6)$$

The effect of the first term in (1.6) is to cause a drift of the light spot at a constant rate across the scale. This can be eliminated by distributing a drift correction over all the readings. If the system is not truly astatic, but has a residual magnetic moment P', the effect of the second term is to cause an error in the deflection

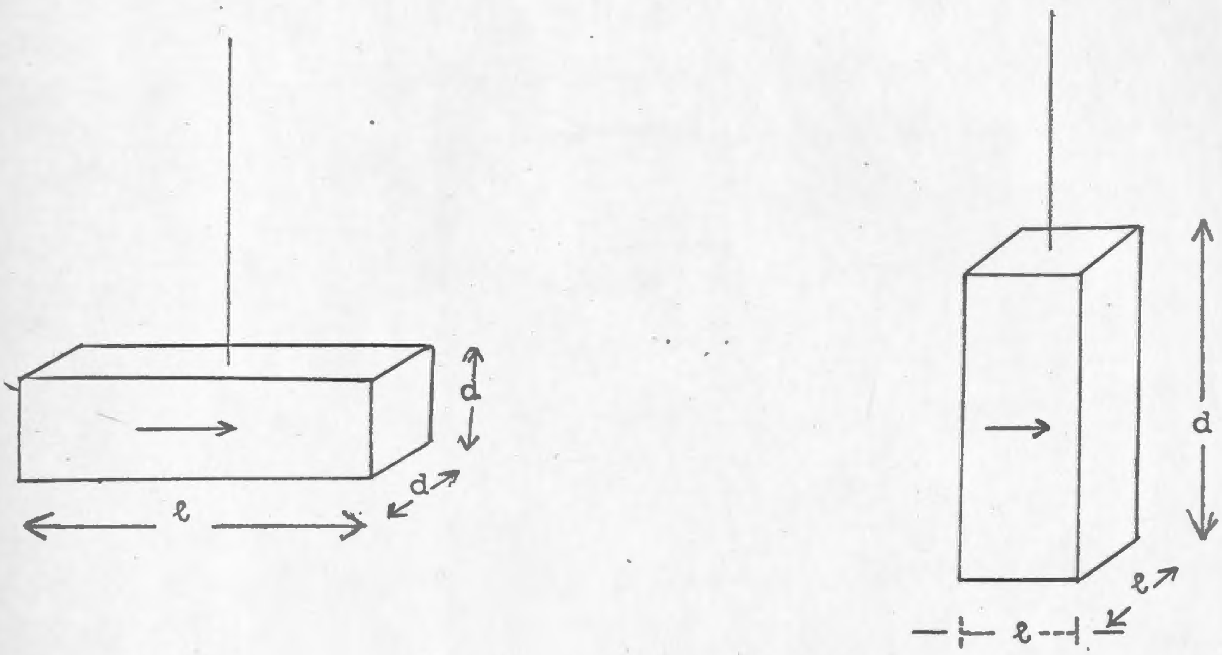
$$\theta' = \frac{P'T^2}{2\sigma} \cdot \frac{d^2H}{dt^2} \qquad (1.7)$$

Blackett has discussed this error and estimated that the maximum value for the period is about 60 seconds in order that the above error may be insignificant.

1.4 Design for maximum value of P/I

As, from the above considerations, the period of the system cannot be increased indefinitely, the design should be made for as large a value of P/I as possible within certain limitations imposed by the frame of the system (see below).

It is general practice to use prismatic magnets of square cross-section in astatic magnetometers. They can be magnetized either (1) parallel to the long side or (2) parallel to one of the short sides. In either case, the magnets are mounted in such a way that their magnetic moments are horizontal as well as antiparallel when the system is suspended (Fig. 2).



$\beta = \ell/d$ = Fineness ratio

(a) $\beta > 1$

$$m = \rho d^2 \ell$$

$$I_o = \frac{\rho \ell d^2}{12}(\ell^2 + d^2)$$

(b) $\beta < 1$

$$m = \rho d \ell^2$$

$$I_o = \frac{\rho d \ell^4}{6}$$

Fig. 2 PRISMATIC MAGNETS MAGNETIZED
(a) LONGITUDINALLY, (b) TRANSVERSELY

The moment of inertia of a transversely magnetized system is then less than that of a longitudinally magnetized system of the same shape and magnetic moment per unit mass, and hence it will be more advantageous to use transversely magnetized magnets. Further, for the same sensitivity, the transversely magnetized system has the advantage over a longitudinally magnetized system of having a shorter period. However, longitudinally magnetized magnets have been largely used in the past because of the unavailability of transverse magnets with sufficiently high specific moments.

There is, however, a realistic lower limit to the moment of inertia of the frame of the system. First, the system should be rigid enough to maintain a high degree of astaticism, and secondly, a mirror which is large enough to give a clear image is to be attached to the frame. Roy (1963) has discussed these limitations in some detail and estimated that a transversely magnetized system with upright magnets will have maximum sensitivity when its total moment of inertia is 4 times the moment of inertia of an individual magnet, i.e., in equation (1.1),

$$I = 4I_0 \qquad (1.8)$$

In selecting the material of the magnets, the following considerations are relevant:

(1) The magnets should have a large dipole moment and a small moment of inertia;

(2) The material of the magnets should have high coercivity to render their magnetic moments insensitive to short-

period relaxation effects or to the action of magnetic torques encountered in the laboratory during their lifetime.

The dipole moment P of a rectangular magnet of mass m is

$$P = m\, J(\beta) \qquad (1.9)$$

where $J(\beta)$ is the dipole moment per unit mass and β is defined as the "fineness ratio", $\beta = l/d$ where l and d are the length and height, respectively, of a magnet of square cross-section (see Fig. 2). Then $J(\beta)$ is a function of the shape of the magnet and may either be calculated from the known B-H curves for the material or found by direct measurement of the dipole moment of the magnet of given mass and shape. Table I gives details of the properties of different permanent magnet materials, while Fig. 3 shows the variation of $J(\beta)$ with fineness ratio β for the same materials. Both Table I and Fig. 3 have been mostly copied from Blackett (1952). However, for the data on Magnadur III, the specifications listed by the manufacturer, Mullard and Co., London, England (1964) have been quoted, and the data for the material Ferroxdure is from Chikazumi (1964). The special advantages of the material Magnadur III will be discussed in Section 1.6.

A practical limit to the useful increase of P is that the field produced over the region of the specimen by the magnets should not be so large as to affect the magnetization of the specimen; i.e., it is necessary to avoid errors in the measure-

TABLE I

PROPERTIES OF SELECTED MATERIALS FOR PERMANENT MAGNETS*

MATERIAL	COMPOSITION	ρ (gm/cm³)	B_r (gauss)	H_c (oe)	$(BH)_{max} \times 10^{-6}$ (gauss x oe)	NO. OF CURVE IN FIGURE 3
Magnadur III	$BaFe_{12}O_{19}$	4.8	3700	3000	3.3	1
Alnico	10Al,17Ni, 12Co,6Cu	7.5	7250	560	1.70	2
Ferroxdure	$BaO,6Fe_2O_3$	4.5	2000	1500	-	3
Alcomax IV	8Al,12Ni,24Co, 6Cu,2Cb	7.5	11200	750	4.3	4
Ticonal K		7.3	9050	1080	3.96	5
Vectolite	$44Fe_3O_4,30Fe_2O_3$ $26Co_2O_3$	3.12	1600	900	0.5	6

ρ = density, B_r = remanence, H_c = coercive force

* Data taken from Blackett (1952), Chikazumi (1964), and Mullard and Co. Catalogue (1964).

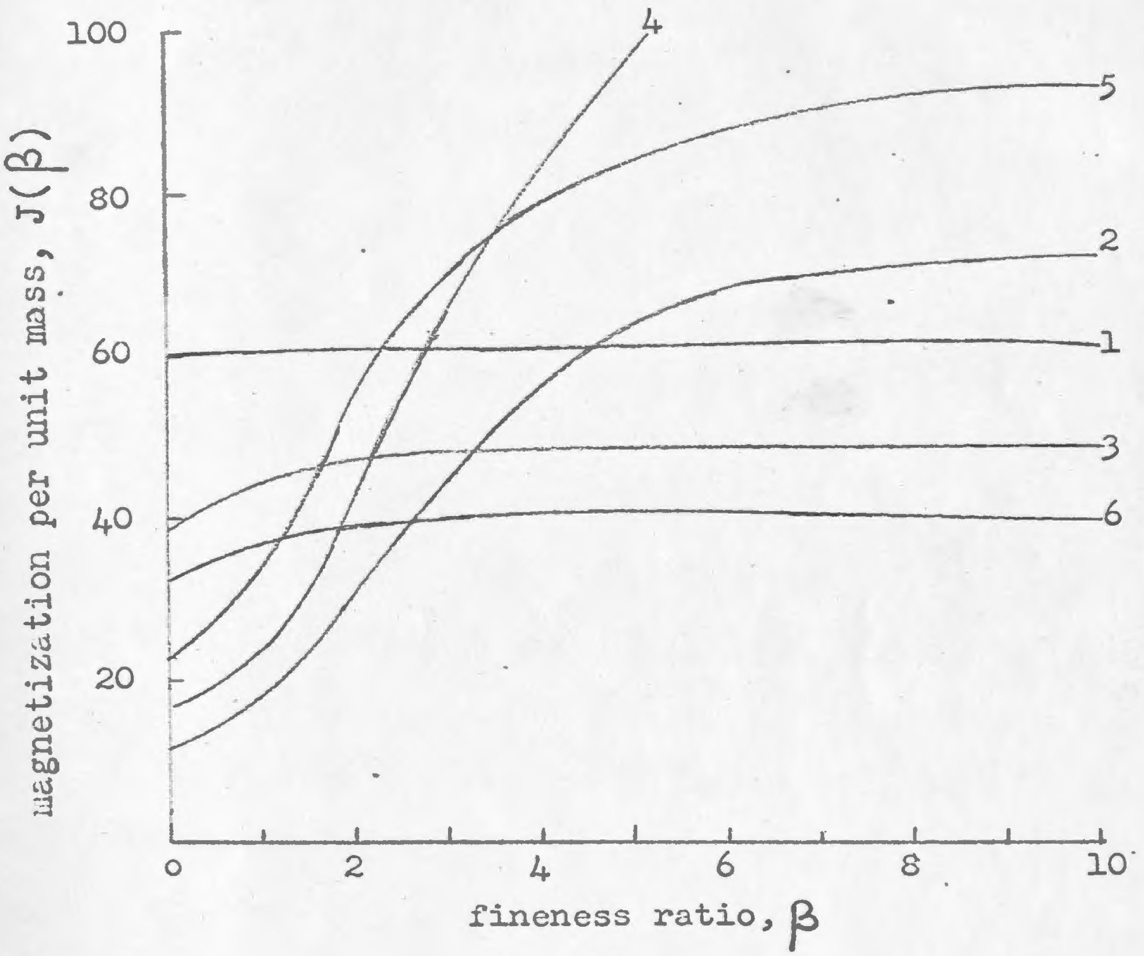


Fig. 3 INTENSITY OF MAGNETIZATION J(β) FOR THE MATERIALS LISTED IN TABLE 1, PLOTTED AGAINST FINENESS RATIO β.

ment of remanence of rock specimens that would be caused if the field due to the astatic system were to introduce a significant viscous component with time constant of the order of the length of a measurement. However, in most practical applications this error can be considered negligible, provided the distance of the specimen from the closest magnet is not too small.

Another parameter to be considered in the design of an astatic magnetometer is the distance L between the two magnets. In general, L should be large enough so that when the specimen is placed close to the lower magnet, the field H_U it produces at the upper magnet is a small fraction of the field H_L at the lower magnet. The usual practice is to vary L between 3 and 20 cms. Even though the above consideration makes it useful to increase L as much as possible, the greater moment of inertia of a long magnetometer and the greater fragility of such a system will offset the advantage. Moreover, it will be difficult to astatize such a long magnetometer, because the production and maintenance of a uniform (zero or finite) field over the region of the system becomes increasingly difficult as its length increases.

1.5 Experimental design of the magnetometer

In the experimental design of the magnetometer, care is taken as much as possible to adhere to the theoretical considerations discussed above.

For the time being, the magnetometer has been installed in a second-floor laboratory of the Science Building, which has serious disadvantages:

(1) The presence of both constant and time-varying fields, caused by ferromagnetic materials and the use of d.c. and a.c. equipment in adjoining parts of the building, can introduce distortions of the magnetic field in the neighbourhood of the astatic system, in the form of permanent field gradients or as more or less temporary disturbances. In the latter case, e.g. as a result of switching on or off the d.c. power in an adjacent room, the null position of the astatic system may be shifted, causing a delay in measurements. Such interference can be avoided through choice of quiet hours for making observations, but permanent gradients are a more serious matter, as they are not removable by the Helmholtz coils surrounding the astatic system (see Section 2.2.) which produce only uniform field components at its center. It was decided, therefore, to conduct a magnetic "survey" of the laboratory, using a compass and dip needle, to locate the space where the astatic system would be surrounded by as uniform a field as possible and temporary electromagnetic disturbances would be at a minimum. As a result of the survey, the magnetometer was assembled in a region near the centre of the room, where the gradient turned out to be insignificant.

(2) Mechanical vibrations transmitted through the floor below the magnetometer resulted in an almost continuous motion of the astatic system, observed at the scale (1.8 meters

distant from the mirror) as a horizontal oscillation of the light spot, with period of 1-2 tenths of a second and a varying amplitude, which, however, was seldom less than 0.5 cm. This effect was eliminated almost entirely by means of a special suspension for the magnetometer housing (Plate 2) which is described in Section 1.8.

1.6 Material of the magnets

In choosing the material of the magnets, advantage was taken of the fact that sintered materials are now available which can be magnetized in a transverse direction. Magnadur III magnets prepared by Mullard Ltd., London, are very suitable for astatic magnetometers as they approach the conditions for optimum design given in Section 1.4. Magnadur III has an unusually high coercivity (3.0×10^3 oe) even for permanent magnet materials and it can be magnetized parallel to one of the short sides with sufficient directional accuracy. Further, it can be seen from Fig. 3 that the value of $J(\beta)$ for Magnadur III is 60 erg/oe, which for $\beta < 1$, is higher than that of any other material mentioned. A total number of twelve transversely magnetized magnets of dimensions $0.75 \times 0.25 \times 0.25$ cm.^{3}, all cut from the same piece, were obtained from the manufacturer, and of these, the two magnets with the closest values of moment of inertia and mass (Table 3) were selected for the present magnetometer. Aluminum was chosen for the material of the frame, to the center of which a circular plane mirror of diameter 4mm. is attached. The two magnets chosen are mounted in opposition, one above the other in the aluminum

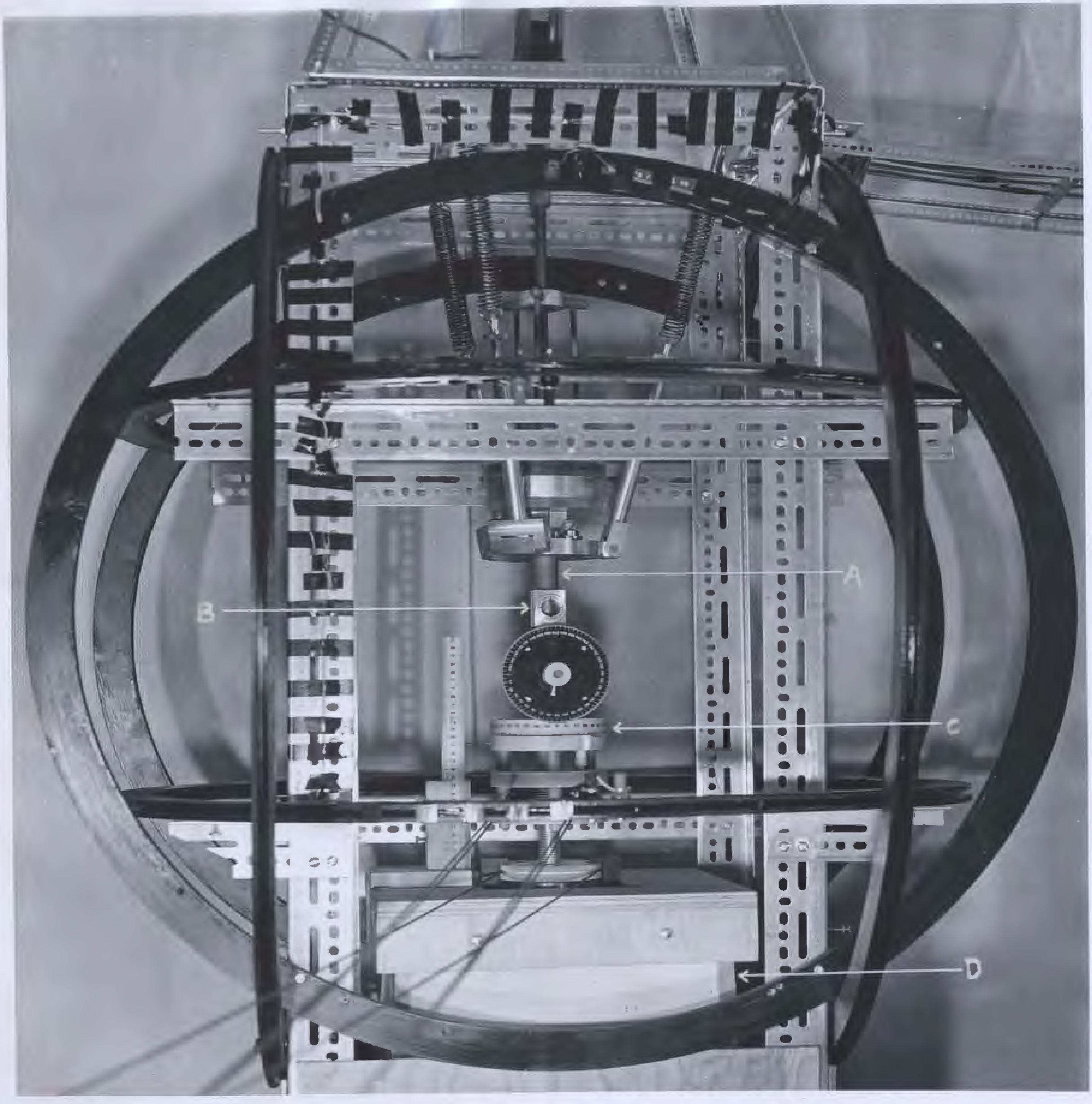


PLATE 1 GENERAL VIEW OF ASTATIC MAGNETOMETER
(FOR SYMBOLS SEE TEXT)

frame and their centres separated by a distance L = 6.0 cms. The final adjustment of the magnets in the frame, and the astatization of the system, are described in Section 2.6.

1.7 Description of the magnetometer

Plate I gives a general view of the instrument. The astatic system is suspended by a thin gold strip of 10 cm. length, and is enclosed in case A made out of tufnol. The case is closed at the bottom by a copper plate to damp the rotational motion of the astatic system by means of the eddy currents induced by the lower magnet. Raising or lowering the damping plate relative to the system permits fine control of the damping. The lens housing B, made of aluminum, contains a viewing window in the front which holds a lens of focal length 1.8 meters. The optical system of the magnetometer consists of the mirror at the centre of the frame, the lens and a lamp-and-scale arrangement at 1.8 meters distance, as illustrated schematically in Fig. I.

1.8 Suspension of the magnetometer housing

A necessary special feature of this magnetometer is the mechanical filter designed to suppress short-period vibrations of impermissibly large amplitude, transmitted to the system through the laboratory floor (Section 1.5). This filter is essentially a top-heavy, elastic anti-vibration support of the system, but using springs under tension for elastic control, rather than thin metal rods under compression as in the Müller suspension used, for example, by Blackett (1952). Three phosphor-

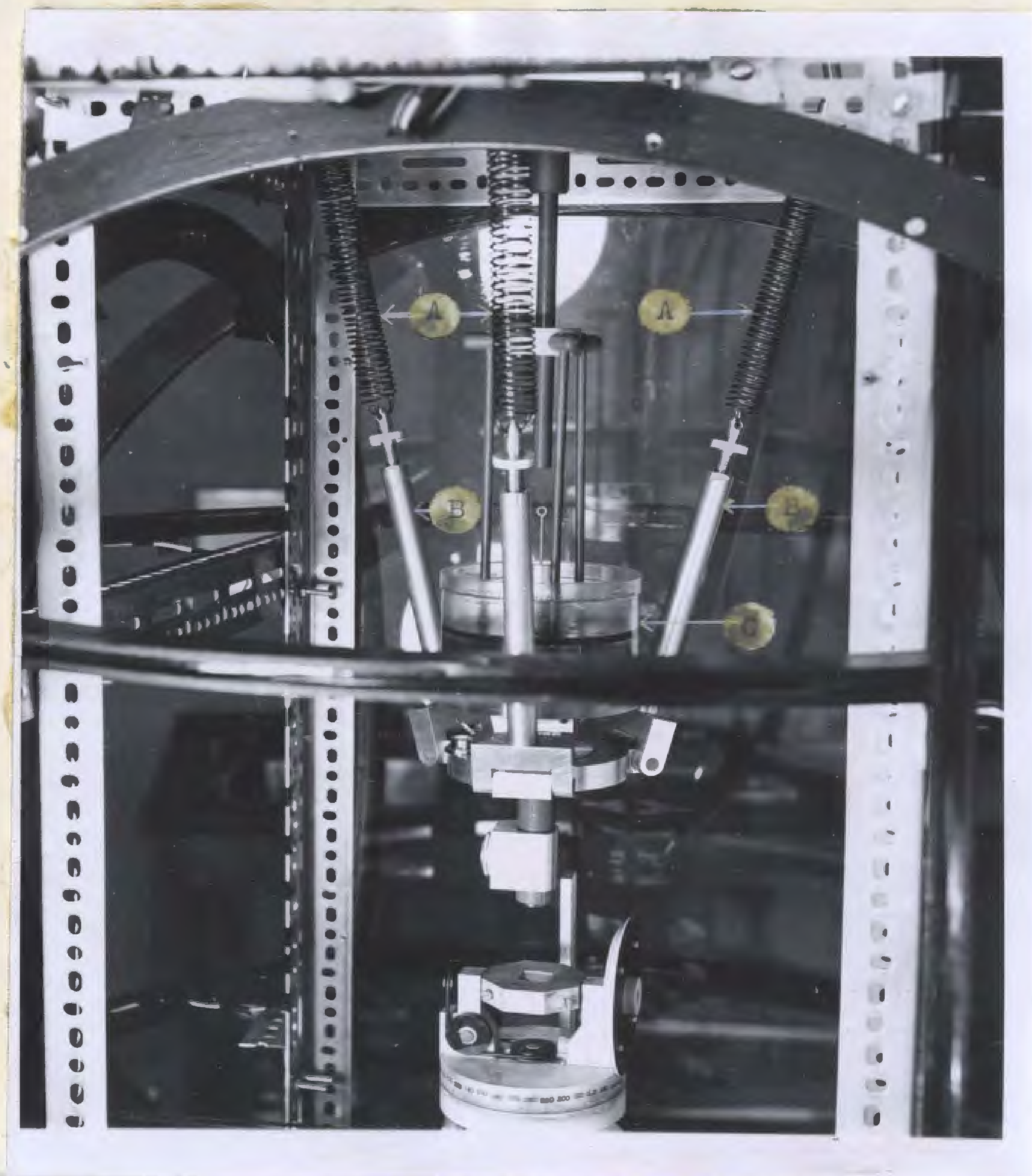


PLATE 2 TOP-HEAVY SUSPENSION FOR THE MAGNETOMETER
HOUSING (FOR SYMBOLS SEE TEXT)

bronze springs (A in Plate 2) were used, each being connected at its lower end to the suspension rods, B, and through these to the base of the magnetometer housing. At the other end the springs are attached to the supporting structure of the Helmholtz coil system, which stands on the floor and is loosely coupled with the ceiling of the room.

The bulk of the top-heavy mass is 2.5 Kg of mercury in a container (C) surrounding and attached to the magnetometer tube. For damping of the natural oscillations of the system, a thin tufnol plate attached to the supporting structure, and adjustable vertically, rests within a layer of light, viscous mineral oil above the mercury.

The dimensions of the springs are as follows:

Diameter of the phosphor-bronze wire:	1.58 mm.
Diameter of a coil of the spring:	26.0 mm.
Number of coils per spring:	40

By means of this arrangement, mechanical vibrations of the light spot, of the magnitude given in Section 1.5, are almost completely eliminated.

The adopted values for the mercury mass and the spring dimensions represent a compromise solution, as it had been hoped to raise the natural period of the finally suspended system to 2 seconds or so. With the existing instrument one could have achieved this only by making the support unrealistically top-heavy or by using springs of very low stiffness. However, the final arrangement, with a natural period of about 0.8 sec., reduced the amplitude of the lateral oscillations of the light spot on the scale to less than 1 mm. at normal times, and to an unobservably low value during quiet times. Therefore the vibration problem may be regarded to have been solved for all practical purposes, though ultimately the best solution will be to mount the magnetometer on a cement foundation in a basement.

1.9 Specimen holder

Plate 3 shows the specimen holder made of laminated fibre plastic. The specimen holder sits on the concrete block D (Plate I) which in turn rests upon the floor of the room, but in isolation from the magnetometer - Helmholtz coil support. The specimen holder can be raised or lowered under the magnetometer, the exact distance from the centre of the lower magnet being measured by means of a cm. - scale mounted vertically at its side. The specimen holder is also capable of being rotated about a vertical or a horizontal axis, with automatic locating stops for every 90° setting. All the operations with the specimen holder could be carried out by remote control, consisting of a pulley-and-wire arrangement, where the operator is seated behind the table supporting the lamp-and-scale system. The specimen can be clamped in the specimen holder in an exact position by means of plastic screws. The exact centering of the specimen holder under the magnetometer is achieved with the aid of a precisely machined pointer attachable to the lower end of the magnetometer tube; in

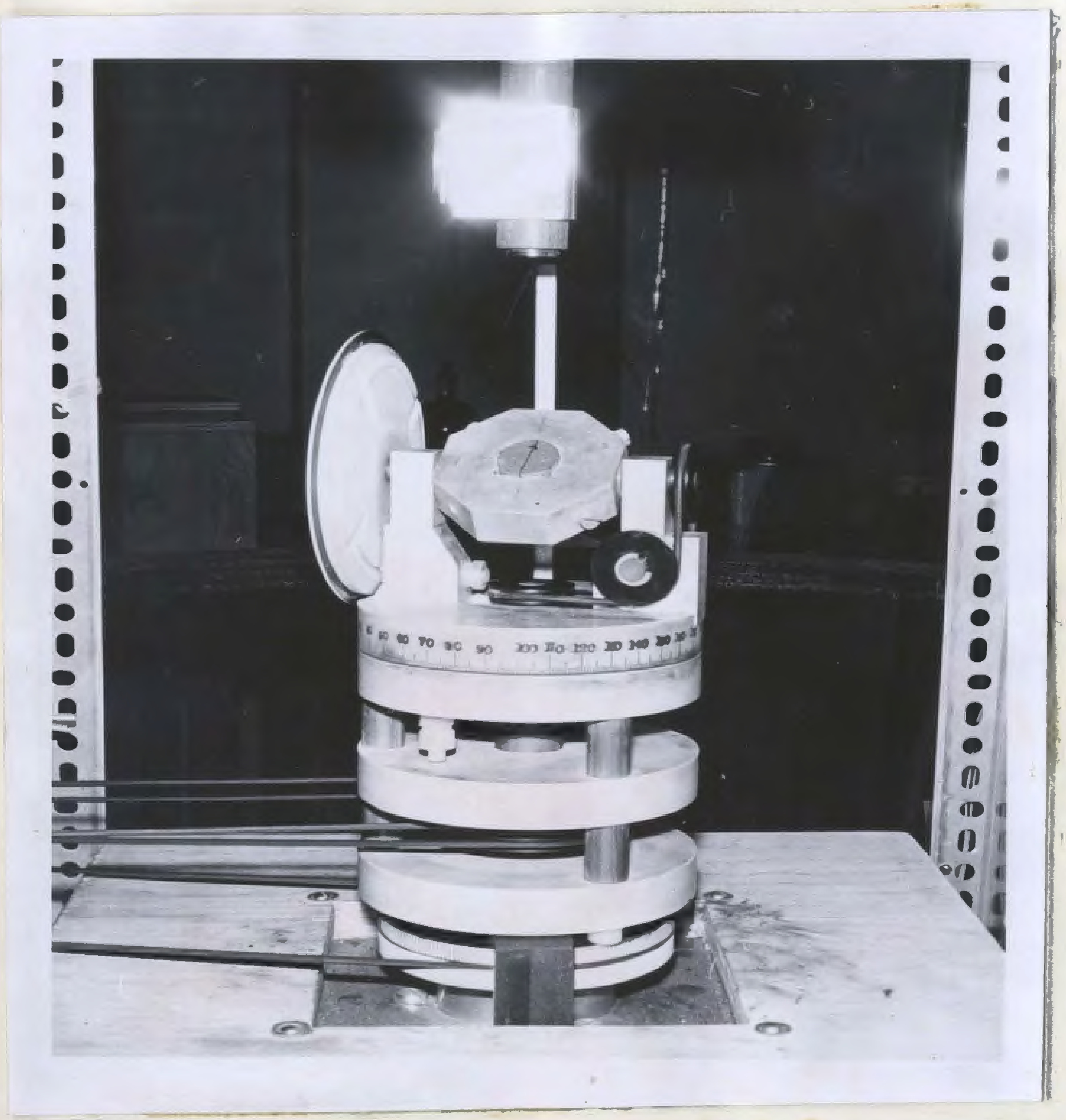


PLATE 3 SPECIMEN HOLDER

this way, the center of the specimen holder can be brought vertically below the axis of the astatic system with an estimated horizontal positioning error of 0.2 mm. or less.

CHAPTER 2

CALIBRATION AND PERFORMANCE OF THE MAGNETOMETER

2.1 Preliminary remarks

The details of calibration and astatization of the astatic magnetometer are reported here. The reciprocal sensitivity of the magnetometer is 3.9×10^{-7} oe per mm. deflection at 1.8 meters scale distance. Following a suggestion by Dr. E. Irving, until recently at the Dominion Observatory, Ottawa, an alternative technique to the use of trimmer magnets was employed as part of the astatization procedure, and this will be discussed in Section 2.6. In Section 2.7, the author's magnetometer is compared with some of the magnetometers described in the literature.

2.2 Helmholtz coil system

One of the fundamental steps in achieving a high sensitivity for the magnetometer is to increase the value of the period T, after assigning the optimum values for the ratio P/I and the torsion constant σ of the suspension fibre (Section 1.5). In a uniform horizontal field H, the period, T, of the astatic system is given by

$$T = 2\pi \sqrt{\frac{I}{\sigma + P'H}} \qquad (2.1)$$

where P' is the horizontal component of the residual magnetic moment of the magnetometer. One of the ways of increasing T is to reduce the magnetic torque P'H compared to the elastic torque σ. Since, in fact, the basic requirement of a good astatic instrument is a low sensitivity to a uniform field, as compared with maximum sensitivity to a field having vertical gradient, any magnetic torques due to a uniform field tending to displace the system from its equilibrium position should be negligible compared with the elastic restoring torque σ. This condition corresponds to

(2.2)

so that when it is fulfilled, (2.1) reduces to:

(2.3)

The magnetic torque can be decreased through reduction of either H or P'. The reduction of P' is otherwise known as "astatizing the system", and will be discussed in Section 2.6. The reduction of H, which is described below, has the further advantage that in a field-free space the induced component of magnetization in the specimen is zero, though, as already noted, care must be taken to ensure that the lower magnet itself does not introduce a significant field in the region of the specimen. Then in the absence of such a field, the remanent magnetization vector, which is of interest in palaeomagnetism, can directly be obtained.

To attain a field-free space, three pairs of circular Helmholtz coils were placed around the magnetometer (E in Plate I) for the compensation of the vertical component, the N-S component, and the E-W component of the geomagnetic field. The E-W coils were placed so that their plane made an angle of about 1° with the magnetic meridian; in this position, a current of about 12 ma is required to compensate the small E-W field. Data relevant to the Helmholtz coils and d.c. power supply are given in Table 2. For a constant field at the magnetometer in the absence of compensation, control of the compensation currents is at present considerably better than the accuracy of measuring the field to be compensated, for which purpose a tangent galvanometer and dip needle have been employed.

The horizontal and vertical field components were measured repeatedly during a period of several weeks in the region of the magnetometer before and after its installation, and while the data did not change significantly with time, the estimated accuracy is only ½% for the horizontal field component and 1% for the vertical component, corresponding to about 100 γ and 500 γ, respectively (1 γ = 10^{-5} oe). In the absence of major fluctuations of the external field, these values also represent the maximum departures from a zero field at the center of the magnetometer. This accuracy was regarded as adequate at the present time, but following the installation of the instrument in a more satisfactory environment, it may be necessary to call for more accurate measurement and control of the compensating field. The fact that the magnetometer at present is operated at a high astatization also reduces the importance of exact compensation of the field components, as the factor P' in the product P'H is already small. Plate 4 shows the arrangement of the control panel for the currents in the Helmholtz coils.

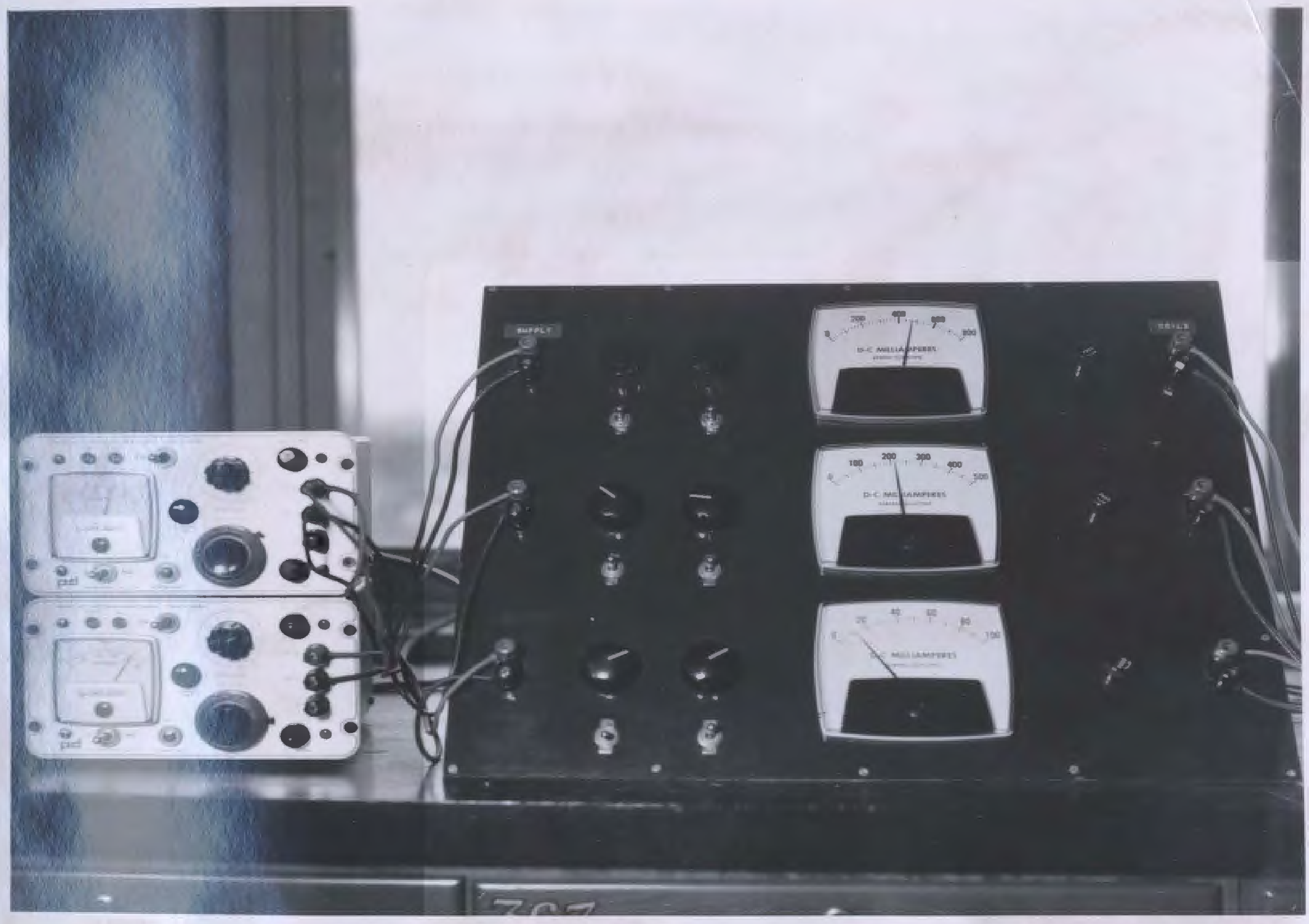


PLATE 4 CONTROL PANEL FOR THE HELMHOLTZ COIL SYSTEM

TABLE 2

DETAILS OF THE HELMHOLTZ COIL SYSTEM FOR COMPENSATING THE GEOMAGNETIC FIELD

1. Dimensions of the Helmholtz coils

Vertical (V) coils	45.4 cm radius 50 turns 4.8 ohms resistance
North-South (N-S) coils	50.2 cm radius 50 turns 5.2 ohms resistance
East-West (E-W) coils	55.1 cm radius 15 turns 1.6 ohms resistance

2. Magnetic field at the centre of the coils

Vertical coils	1.00 oe/amp
North-South coils	0.91 oe/amp
East-West coils	0.25 oe/amp

3. Components of the geomagnetic field

Vertical component, H_V	0.466 oe $\pm .005$
Horizontal component, H_H	0.194 oe $\pm .001$
Inclination, I	67.5° North downward

4. Currents needed in the coils for compensation of the above components

Vertical coils	475 mA
North-South coils	215 mA
East-West coils	12 mA

TABLE 2 (CONTINUED)

5. Total error that might occur in the compensation

Vertical coils	1.0%
North-South coils	1.0%
East-West coils	3.0%

6. Power Supply

Model used:	Precision Power Source Model 605, (Power Designs, Inc.).
Stability:	100 micro-volts for eight-hour period, with fixed load and constant a.c. line voltage.

2.3 Suspension of the astatic system

The rectangular frame of the system has an extension on top, to which is attached a gold strip of length 10 cm, giving a torsional constant $\sigma = 1.41 \times 10^{-2}$ dyne-cm/radian. The system is suspended such that the north pole of the lower magnet points north. The details of the magnetometer are tabulated in Tables 3 and 5.

2.4 Sensitivity of the magnetometer

The theoretical sensitivity of the astatic magnetometer has been given as

$$\frac{\theta}{H} = \frac{T^2}{4\pi^2} \cdot \frac{P}{I} \qquad (1.5)$$

with symbols and units as before. θ/H can be found after substitution of experimentally determined values of P, I, and T into (1.5), where

$$P = 10.1 \text{ erg/oe},$$
$$I = 7.5 \times 10^{-3} \text{gm-cm}^2,$$
$$\text{and} \quad T = 4.6 \text{ secs.}$$

(The same value, T = 4.6 secs., is also obtained when the experimentally determined values of I, as above, and $\sigma = 1.41 \times 10^{-2}$ dyne-cm/rad. are substituted into equation 2.3).

Then $\theta/H = 726$ rad./oe

and, converting from radians to mm. deflection at a scale distance

TABLE 3

DETAILS OF THE ASTATIC MAGNETOMETER

1. MAGNETS

Mass, m(grams)	Upper magnet:	0.2116
	Lower magnet:	0.2117
Dipole moment of either magnet, P(ergs/oersted):		10.1
Moment of inertia of either magnet, I_0 ($gm\text{-}cm^2$):		0.0022
Height of either magnet, h (cms):		0.75
Width and breadth of either magnet, w = b (cms):		0.25 x 0.25
Side/height ratio, $\beta = w/h = b/h$:		0.33
Material (ceramic):		'Magnadur 3' (powdered barium ferrite)
Magnetization direction:		Transverse (parallel to a short side)
Coercivity, H_C (oersteds)		3.0×10^3
Maximum flux density x demagnetising field, $(BH)_{max}$ (gauss-oe)		3.6×10^6

2. ASTATIC SYSTEM

Material of the magnet holder:	Aluminum
Moment of inertia of magnet holder and mirror, I_1 ($gm\text{-}cm^2$):	0.0031
Total moment of inertia of astatic system, $I_1 + 2I_0$ ($gm\text{-}cm^2$):	0.0075
Moment of inertia ratio, $= (I_1+2I_0)/I_0$:	3.4
Separation of the magnet centres, L(cms):	6.0
Polarity:	North-seeking end of lower magnet pointing North

TABLE 3 (CONTINUED)

3. SUSPENSION

Length, (cms):	10.0
Cross-section (width x breadth), w' x b' (cms x cms):	0.005 x 0.0005
Torsional constant, (dyne-cm/radian):	0.014
Material:	Gold

of 1.8 meters, the sensitivity can be written as

$$\theta/H = 2.62 \times 10^{6} \text{mm./oe}$$

It is customary to express the performance of the magnetometer in terms of its reciprocal sensitivity, H/θ. Thus the theoretical reciprocal sensitivity (from measured constants of the magnetometer) is

$H/\theta = 3.82 \times 10^{-7}$ oe per mm. deflection at 1.8 meters scale distance.

2.5 Calibration of the magnetometer

The astatic magnetometer was calibrated as described below. A small circular coil of n turns and cross-sectional area A was placed vertically above the magnetometer. The distance of the coil above the magnetometer was kept much greater than the separation between the magnets, so that the vertical gradient of the horizontal field produced by the coil will be nearly constant at the magnetometer. The calibration coil was oriented with its plane vertical and its axis at right angles to the magnetic moments of the magnets. By passing a small current through the coil, an equivalent magnet of known horizontal magnetic moment, Me, was produced. This is given by

$$Me = \frac{nAi}{10} \text{ erg/oe} \qquad (2.4)$$

where A is in cm^2

and i is in amperes

This produces a constant vertical gradient dH/dZ at the magnetometer, which may be expressed in terms of the difference in magnetic intensity, ΔH, at the two magnets, v given by

$$\Delta H = L \frac{dH}{dZ} = M_e \left[\frac{1}{Z_u^3} - \frac{1}{Z_L^3} \right]$$

$$= -3L \frac{M_e}{\bar{Z}^4} \qquad (2.5)$$

where Z_u, Z_L are the distances from the centre of the calibration coil to the centres of the upper and lower magnet, respectively, L is the distance between the magnets,

and $\bar{Z} = Z_u + \frac{L}{2}$

If this field acting on the magnetometer produces a deflection D(mm.) on a scale placed at a certain distance from the magnetometer mirror, the sensitivity may be defined as

$$S = D/\Delta H \quad \text{mm. per oe} \qquad (2.6)$$

The sensitivity thus defined depends on the distance of the lamp and scale from the magnetometer. Then the reciprocal sensitivity in the present case may be expressed as

$$1/S = \Delta H/D \quad \text{oe. per mm.} \qquad (2.7)$$

and substituting (2.4) and (2.5) into (2.7),

$$\frac{1}{S} = \frac{nAi}{10D} \left(\frac{1}{Z_u^3} - \frac{1}{Z_L^3} \right)$$

oe per mm. deflection at 1.8 meters scale distance (2.8)

The deflection D was measured for two values of $\overline{Z}$, and the observations, which agreed to less than 1% are tabulated in Table 4.

Following Blackett's (1952) notation, the gradient sensitivity may be defined as:

" 1 mm. deflection corresponds to g' oe " and expressed as

$$g' = 3.88 \times 10^{-7} \text{oe (at 1.8 m. scale distance)}$$

This is the observed value, which agrees well with the value of $\theta/H = 3.82 \times 10^{-7}$oe (at 1.8 m. scale distance) for the theoretical reciprocal sensitivity (Section 2.4).

2.6 Astatization of the magnetometer

It is not possible to make the dipole moments of the two magnets of the magnetometer exactly equal and opposite. In practice, a resultant dipole moment P' will always remain, so that the suspended system will respond to a changing uniform field as if it were a single magnet of horizontal moment P'. The astaticism of the magnetometer is defined as the ratio of the individual dipole moment P to the residual dipole moment P', i.e.

$$S = P/P' \qquad (2.9)$$

The process of astatization consists of making the resultant dipole moment P' as small as possible.

TABLE 4

CALIBRATION OF THE ASTATIC MAGNETOMETER

DETAILS OF THE CALIBRATION COIL

Number of turns, n:	56
Radius, r (cms):	2.25
Current passed, I (amps):	1.00
Equivalent magnetic moment of the coil, M_e (ergs/oe):	89.0
Distance from lamp-and-scale unit to center of astatic system (meters):	1.78

Case I

Mean distance $\overline{Z}$ of the coil from the magnetometer (cms):	127.8
Field on the magnetometer due to the coil, H(oersteds):	6.02×10^{-6}
Deflection observed on the scale, D_1(mms):	15.5
Reciprocal sensitivity, g', (oe per mm deflection):	3.88×10^{-7}

Case 2

Mean distance $\overline{Z}$ of the coil from the magnetometer (cms):	71.4
Field on the magnetometer due to the coil, H(oersteds):	6.17×10^{-5}
Deflection observed on the scale, D_2(mms):	158.0
Reciprocal sensitivity, g', (oe per mm deflection):	3.91×10^{-7}

TABLE 4 (CONTINUED)

Mean Reciprocal Sensitivity, g'

Mean value of g' at 1.78 meters scale distance: (oe/mm deflection):	3.90×10^{-7}
Theoretical value of g' at same scale distance (Equation 1.5 with measured values of T and P/I):	3.82×10^{-7}

The moments of the two magnets are made as nearly equal in magnitude as possible before they are mounted in the frame. All 12 magnets had been machined to close similarity in shape, and the pair finally selected differed by less than 0.05% in mass and in magnetic moment to begin with. However, the small difference in P was further reduced by rubbing the stronger magnet with emery paper until no difference between the moments of the two magnets could be detected. An alternative method would have been to apply a decreasing a.c. field to produce the necessary slight demagnetization in the stronger magnet. The amount of astatization obtained at this stage can be estimated by comparing the period of oscillation of the system, suspended by a non-elastic fibre, when the magnets are mounted nearly antiparallel (T_1) and parallel (T_2), respectively:

Then $$S = P/P' = \frac{1}{2}\,\frac{T_1^2}{T_2^2} \qquad (2.10)$$

An astatization of S = 100 to several hundred can be obtained quite easily by this method. In the present case a value of S = 400, approximately, had been reached at this stage of the astatization process, though the method of measuring S was not as given by (2.10) which becomes increasingly difficult to use in view of the lengthening of the period T as S increases. Instead, the method of measuring S outlined below was used during the entire astatization process.

For attaining higher values of S, Blackett (1952) suggested the use of "trimming" magnets, or "trimmers" made in the form of short lengths of wire from an alloy with good ferromagnetic properties. The trimmers are attached to the aluminum frame by means of horizontal pivots which permit their rotation in a vertical plane. It is customary to use two trimmers with their pivot axes perpendicular to one another and nearly parallel and perpendicular to the magnetic moment vectors of the main magnets, respectively. Through adjustment of the angles which the trimmers make with the vertical axis of the magnetometer, the magnitudes of the horizontal components of their moments can be made almost equal and opposite to the corresponding residual moments in the two horizontal directions, due to the main magnets. In this way, a high astatization may be obtained. Blackett (1952) reported that in his Vectolite magnetometer (Table 5) values of S as large as 5000 could be maintained for many weeks, and considerably larger values for short periods, but he set an upper limit of S = 10,000 as the highest value he was able to measure reliably.

The use of trimmer magnets, though successfully employed by subsequent research workers (e.g. Watkins, 1961), has the disadvantage that it will increase the moment of inertia of the system.

In the present set-up the method suggested by Irving (Section 2.1) was used to complete the astatization of the system after matching of the magnets and treatment with emery paper had produced a value of S = 400 or so when they were mounted anti-parallel in the holder. A third Magnadur III magnet was broken into small chips and as all chips in turn had the properties of stable dipoles, they could be positioned in such a way as to oppose the residual vector components due to the main magnets, and hence used in astatizing the system. In this method the chips are placed directly upon a surface of one of the main magnets, and turned into desired positions with a small non-magnetic needle. The procedure outlined below was followed:

The magnetometer was suspended by the gold-strip of required length (10 cms) at the centre of the compensating coil system. A uniform horizontal field was applied by passing a small current (in excess of the field compensation current) through the N-S Helmholtz coils. The magnetometer system will be deflected because of the out-of-balance moment P' due to imperfect astatization. Again following Blackett's (1952) notation, this sensitivity to a uniform field can be expressed as:

"1mm deflection corresponds to f_1 oe".

f_1 was now measured for various orientations of a magnetic chip placed on the surface of the lower magnet. In rotating this chip, it was not necessary for its magnetic moment vector to remain in the same plane after each adjustment, but only that successive adjustments should produce a systematic approach to the condition where the horizontal component of its moment would be equal and opposite to the out-of-balance moment P' due to the main magnets. The greater maneuverability of the chips compared to that of pivoted trimmers thus offers some advantage, though in practice two or more chips, with their moments aligned in different planes, are required to achieve the best astatization.

After each adjustment of the first chip, the value of f_1 was remeasured until it proved difficult to increase f_1 any further.

A small current in excess of the field compensation current was then passed through the E-W Helmholtz coils, and the field sensitivity f_2, again corresponding to a scale deflection of 1mm, was measured. The orientation of the magnetic particle on the surface of the magnet was adjusted until f_2 could not easily be increased further, but without reducing f_1 below its previous maximum value. This required careful manipulation, including the use of a second chip, placed upon the surface of the upper magnet. When f_1 and f_2 had been made as large as possible, the particles were affixed on the respective magnet surfaces with a drop of cement each.

This procedure of astatizing the system has the decided advantage over the technique employing trimmers, in that the moment of inertia of the system will be increased by a much smaller amount through addition of the chips. Moreover, an astatization once obtained tends to be more durable, as the chips are derived from the same material as the main magnets, and with the possible exception of some differences arising from the shape factors, will have the same properties, including the exceptionally high coercivity of Magnadur III. These advantages probably offset the disadvantage caused by a relative lack of geometrical control in adjusting the positions of the chips, as compared with the adjustment of trimmers which is often controllable to 1^{o}.

As mentioned previously, the astatic magnetometer, from its principle, will have a high sensitivity to a non-uniform magnetic field and a low sensitivity to a uniform field.

Then following Blackett, the astatization of the magnetometer may be expressed in terms of two components S_1 and S_2 in the N-S and the E-W directions, respectively. They may be written as:

$$S_1 = \frac{\text{N-S uniform field sensitivity (reciprocal)}}{\text{non-uniform field sensitivity (reciprocal)}} = \frac{f_1}{g'} \qquad (2.11)$$

$$S_2 = \frac{\text{E-W uniform field sensitivity (reciprocal)}}{\text{non-uniform field sensitivity (reciprocal)}} \qquad (2.12)$$

A current of 100ma, equivalent to a field H = 0.091oe. directed southward at the centre of the coil system, was passed through the N-S coils in excess of the compensation current. The resulting scale deflection was 36mm at a scale distance of 1.8 meters. Hence, from (2.11),

$$S = \frac{f_1}{g'} = \frac{0.091}{36} \times \frac{10}{3.9} = 6.5 \times 10^3$$

where the experimental value of g' from two measurements listed in Table 4 was used.

A current of 80ma, corresponding to H = 0.020oe. directed westward at the centre of the coil system was then passed through the E-W coils in excess of the compensation current. This resulted in a deflection of 6.0 mms, on the scale. Hence

$$S_2 = \frac{f_2}{g'} = \frac{0.020}{6.0} \times \frac{10^7}{3.9} = 8.0 \times 10^3$$

The procedure followed here therefore led to the attainment of a relatively high astatization, with a value of S in excess of 6000. This is a satisfactory result, as a high astaticism is necessary so that the system can be made relatively immune to changes and fluctuations in uniform external fields. While this makes it possible to employ correspondingly longer periods of oscillation of the astatic system, and thus to render it more sensitive to gradient fields, the relatively low period of T = 4.6 secs. was adopted for the reasons already discussed.

2.7 Comparison with other magnetometers

Table 5 gives a comparison of the present magnetometer with instruments constructed at Manchester, Ottawa and Newcastle upon Tyne; and described by Blackett (1952), Roy (1963), and Collinson and Creer (1960), respectively.

The following are some salient features of the present

TABLE 5

COMPARISON OF DIFFERENT MAGNETOMETERS

Author	Present	Blackett (1952)	Blackett (1952)	Roy (1963)	Collinson & Creer (1960)
Material of magnets:	Magnadur 3	Vectolite	Alcomax IV	Ferroxdure	Alcomax IV
Coercivity (oe):	3000	900	750	1500	750
Magnetization direction:	Transverse	Transverse	Longitudinal	Transverse	Longitudinal
Dipole moment P(ergs/oe):	10.1	60.0	15.0	7.5	2.3
Moment of inertia of magnet system, I(gm-cm^2):	0.0075	0.165	0.041	0.017	0.0012
Dimensions of magnets, (cm x cm x cm):	0.75 x 0.25 x 0.25	3.2 x 0.5 x 0.5	0.80 x 0.18 x 0.18	0.74 x 0.30 x 0.30	0.30 x 0.15 x 0.15
Fineness ratio, β	0.33	0.16	4.5	0.40	2.0
$\alpha = I/I_0$:	3.4	4.5	4.1	3.5	2.5
P/I:	1350	364	366	440	1920
Torsional constant of suspension (dyne cm/rad):	141×10^{-2}	104×10^{-2}	1.80×10^{-3}	1.76×10^{-3}	2.41×10^{-2}
Length of suspension (cms):	10.0	-	-	10.6	6.0

TABLE 5 (CONTINUED)

Author	Present	Blackett (1952)	Blackett (1952)	Roy (1963)	Collinson & Creer (1960)
Period T(secs):	4.6	25	30	19.5	1.4
Theoretical sensitivity (rad/oe):	726	5760	8350	4250	114
Lamp-and-scale distance from the magnetometer (meters):	1.8	5.0	5.0	2.5	1.0
Field for 1mm deflection at the above lamp-and-scale distance (oe per mm deflection)	3.9×10^{-7}	1.2×10^{-8}	1.5×10^{-8}	4.7×10^{-8}	2.0×10^{-7}

magnetometer:

(1) The coercivity of the material Magnadur III is higher than that of the three materials used in the four other magnetometers.

(2) It can be seen from Fig. 3, that, for Magnadur III, the value of the shape-dependent specific magnetization $J(\beta)$, (60 c.g.s. units) is larger than that of the other three materials in the region $\beta<1$. Only the magnets in Blackett's Alcomax IV magnetometer had larger $J(\beta)$, but this is at the expense of an increase in moment of inertia.

(3) The high value of P/I achieved, enables the system to be operated at a comparatively smaller period for a given sensitivity. The period (T = 4.6 secs.) is substantially less than that of the magnetometers by Blackett and Roy, thus permitting more rapid measurements, but Collinson and Creer's instrument with P/I = 1,920 and T = 1.4 secs., surpasses all others, including the author's, in this respect.

(4) High value of astaticism was achieved by a relatively easy method.

(5) The special suspension system of the magnetometer housing allow the instrument to be operated in the presence of otherwise intolerable mechanical vibrations of the laboratory floor.

The actual sensitivity ($1/g'$) of the author's magnetometer is the lowest among those listed in Table 5, but as discussed below, an increase in $1/g'$ by as much as a factor of 10 can be achieved without too much difficulty.

The relatively low sensitivity of the author's magnetometer, results mainly from the use of a coarse suspension fibre (Tables 3 and 4). If a finer fibre (similar to that employed by Roy, with $\sigma = 1.76 \times 10^{-3}$ dyne-cm/rad) had been used as a suspension of the author's astatic system, the period and the reciprocal sensitivity would have been T = 13s and g' = 4.8×10^{-8} oe per mm deflection, respectively, i.e., the sensitivity could have been improved by about an order of magnitude by using the finer suspension fibre. As an alternative, the sensitivity can also be improved by increasing the length of the fibre, as σ is proportional to its length. A further limited increase in sensitivity may be achieved by increasing the lamp and scale distance from the magnetometer but only at the sacrifice of some optical properties.

In any case, the sensitivity of the present magnetometer was found to be sufficient for the measurement of directions of even weakly magnetized sedimentary rocks, with intensities of magnetization as low as, say 1×10^{-5} emu/cm^3. Hence no attempt was made either to lengthen the fibre or change the suspension to finer gold or phosphor-bronze ribbon or a quartz fibre. Moreover, as the previously mentioned presence of disturbing fields and mechanical vibrations in the neighbourhood of the present location of the instrument sets a practical upper limit to its useful sensitivity, it was decided not to improve the sensitivity any further until the magnetometer has been installed in better surroundings.

2.8 Accuracy of observations

The r.m.s. deflection θ_0 of the astatic system due to thermal noise is given by:

$$\frac{1}{2}\sigma\theta_0^2 = \frac{\epsilon}{2} \qquad (2.13)$$

where $\epsilon = Kt$, K is Boltzmann's constant, and t is the temperature in degrees absolute.

On substituting (1.3) into this equation:

$$\theta_0 = \frac{\epsilon^{1/2}T}{2\pi I^{1/2}} \qquad (2.14)$$

Blackett (1952) introduced a quantity H_0, defined as the magnetic field which will produce a deflection equal to the r.m.s. thermal deflection θ_0 .

Then

$$PH_0 = \sigma\theta_0$$

and

$$H_0 = \frac{2\pi\epsilon^{1/2}}{T} \cdot \frac{I^{1/2}}{P} \qquad (2.15)$$

The quantity H_0, from its definition, is the standard error of a single reading of the magnetic field due to the thermal motion of the suspended system. H_0 is called the minimum detectable field of the astatic magnetometer. This from the constants of the magnetometer, the minimum detectable field of the present instrument is 3.34×10^{-9} oe.

The standard error δ_0 due to the thermal oscillation of the suspended system is given by:

$$\delta_0 = 2 R \theta_0 \qquad (2.16)$$

where R is the distance of the scale from the magnetometer mirror.

It can be expressed as:

$$\delta_0 = \left(\frac{2 R \epsilon}{g' P} \right)^{1/2} \qquad (2.17)$$

where δ_0 and R are measured in mm. When the known values for R, E, g' and P are inserted into (2.17), the computed thermal deflection δ_0 of the light spot on the scale (at 1.8 m. distance from the magnetometer) becomes:

$$\delta_0 = 6 \text{ microns.}$$

In practice many other small torques act upon the magnetometer system and thus contribute to its noise level, in addition to thermal fluctuation torques. It is fairly difficult to estimate the standard error due to combined noise parameters.

However, a more realistic approach is to calculate the standard deviation from the mean of a series of observations for a standard signal. This will be discussed later in Section 3.6.

CHAPTER 3

MEASUREMENTS IN PALAEOMAGNETISM USING AN ASTATIC MAGNETOMETER

3.1 Preliminary remarks

The astatic magnetometer is a basic instrument to measure the direction of natural remanent magnetization of rocks. One can compute the total magnetic vector by measuring the various components of magnetization. The contents of this chapter are concerned with the method of making such measurements with an astatic magnetometer. Later sections of the chapter deal with the processing of the palaeomagnetic results, the procedure of determining the mean direction of magnetization of the formation from the specimen data, and a discussion of errors.

3.2 General considerations

The following convention was followed with regard to notation: An irregular rock piece collected from the field, and with the required orientation marks, is referred to as "sample". A rock piece cut from the above sample in the form of a regular geometric shape, and which is ready to be measured, is referred to as "specimen".

The magnetization of a homogeneously magnetized specimen can be approximated to that of a dipole at its centre, provided that the distance of the specimen from the lower magnet exceeds a certain minimum value (see below). If the specimen is placed under the magnetometer, the horizontal component of its dipole will produce a gradient field, causing the astatic system to be deflected. By noting the polarity of the lower magnet and by positioning the rock specimen in different orientations under the magnetometer, the direction of the magnetic vector in the rock can be computed.

There are three positions relative to the magnet system in which the specimens may be placed for measurement:

(a) Directly under the lower magnet of the astatic system;

(b) To one side of the lower magnet;

(c) To one side of the astatic system, midway between the magnets.

Collinson and Creer (1960) have discussed the relative merits of the three positions and have shown that positions (a) and (b) are more advantageous than position (c). The main advantage of position (b) over (a) is that it permits the magnetic dip (defined as the angle which the dipole makes with the horizontal plane and taken as positive when the north pole points downward) to be calculated from measurements obtained when the specimen has been displaced sideways along a line perpendicular to the plane containing the magnetic moments of the astatic system. This avoids the more awkward procedure of turning cylindrical specimens about a horizontal axis during dip measurements, which means, moreover, that the specimen center cannot be brought as close to the lower magnet as is possible during measurements of magnetic azimuth (i.e. the horizontal component), which involve rotations about the vertical axis below the astatic system. Because of this, position (b) is particularly useful with weakly magnetized specimens in the form of thin discs.

However, this advantage is offset by disadvantages. Papapetrou (quoted in Blackett, 1952) showed that the remanence of a homogeneously magnetized cylindrical specimen can be represented by horizontal and vertical dipole components P_x, P_z, respectively, at its center, and he derived expressions for the horizontal deflecting fields H_x and H_z at the lower magnet of the magnetometer due to P_x and P_z, respectively. Then if z is the vertical distance of the specimen center below the center of the lower magnet, x is the sideways displacement, and if x/z is small and the effect of the upper magnet is assumed to be negligible (i.e. L is large),

$$H_x = P_x F_x / z^3 \qquad H_z = 3P_z x F_z / z^4 \qquad (3.1)$$

where F_x and F_z are functions of z and the specimen shape, and represent the factors by which the field of the finite cylinder deviates from that due to a centered dipole. For specimens of radius a and height h, Papapetrou then estimated the deviation from the dipole assumption as a function of z and specimen shape. His calculations show that the dipole deviation factors F_x and F_z not only approach unity as a specimen of given radius increases its vertical distance from the lower magnet (i.e. as z/a increases), but that these factors become negligible for much lower minimum values of z/a when the specimen is relatively tall (i.e. h/a is large) than for thin discs. This is shown by the following data selected from Papapetrou's table:

z/a	h = 2a		h = 0.4a	
	F_x	F_z	F_x	F_z
1.5	.815	.535	.555	.400
3.0	1.02	1.02	.862	.774
7.5	1.01	1.01	.974	.948
15.0	1.00	1.00	.994	.989

Moreover, it follows from equation (3.1) that the dip calculation depends on the ratio F_x/F_z, which equals 1.00 for z/a = 3.0 in the case of the "square" cylinder (h = 2a), so that the error in the dip measurement is already negligible for this low value of z/a. In the case of the disc (h = 0.4a) the condition $F_x/F_z < 1.01$ is only reached at $z/a > 15$.

It follows that in the absence of an awkward calibration, flat specimens have to be measured at larger vertical distances from the lower magnet than specimens of the same radius approaching a square cross-section. This removes the advantage of position (b) which would otherwise allow weakly magnetized discs to remain close to the lower magnet even when they are offset during dip measurement.

For the above reason, it was decided to adopt position (a), and the specimen holder was constructed accordingly (Plate 3). While different authors favour different values of h/a when using cylindrical specimens, the adopted values are largely in the range h/a = 1.75 to 2.00. The latter ratio has been chosen here for the cylindrical specimens, which are sliced from cores obtained in the laboratory with a diamond drill of 7/8 inch (2.2 cm) internal diameter; i.e. the diameter as well as the height of the specimen is 2.2 cms.

In order to approach the dipole approximation as closely as possible, and also to prevent significant induction effects in the specimen due to the lower magnet, specimens should be placed as far below the astatic system as is consistent with accurate measurement, and this rule has been followed. Another reason applies when, as sometimes happens, the rock specimen is inhomogeneously magnetized. Provided that z is relatively large, Collinson, Creer, Irving, and Runcorn (1957) have shown that inhomogeneity can be represented by a small displacement of the dipole within the specimen and can be averaged out by taking sufficient readings with the specimen in different positions.

3.3 Procedure for measuring the direction of magnetization of rock specimens

The direction of magnetization of a rock specimen is usually represented in terms of two components: (1) the horizontal component or azimuth of the magnetic vector, and (2) the dip. A knowledge of the azimuth and dip of the magnetic vector gives complete data about the direction of magnetization. Most of the palaeomagnetic investigations reported in this thesis were carried out on cylindrical specimens, but specimens from the Virgin Rock shoals in the Grand Banks were cut in the form of cubes, and a different procedure was adopted for measuring their magnetization. Measurements with the cubes will be reported in Chapter 6.

Normally, the rocks have been oriented in the field with respect to the horizontal plane and to a horizontal direction which is usually magnetic or true north; in the former case, the true north direction is easily substituted by correcting for the local magnetic declination. In actual practice, the angle between the magnetic or true north and a horizontal "bearing" line marked on the rock in its in-situ position is measured in the field. Usually it is found convenient to make the axis of a core drilled from the oriented sample parallel with the in-situ vertical direction, and generally two or more cylinders are sliced from each core. The bearing direction is then transferred to the flat surfaces of each specimen in the form of diagonal arrows.

The first operation on the cylindrical specimens is the measurement of the azimuth of the magnetic vector, which is made with respect to the bearing direction. When the azimuth has been determined, its direction is marked on the cylinder surfaces, again in the form of diagonal arrows. This defines the vertical plane of the magnetization vector, and the specimen is then turned in azimuth, so that this plane is at right angles to the magnetic moments of the magnets in the astatic system. In this position, the specimen is then turned about a horizontal axis for the measurement of magnetic dip. The measurement of azimuth and dip in cylindrical specimens is now a standard procedure, and details will not be given; these can be obtained from a number of sources, e.g. Watkins (1961).

3.4 Calculation of the intensity of magnetization

The intensity of magnetization J of the specimen can be obtained by direct integration over the whole volume of the specimen. This yields the following formula:

$$J = \frac{1}{SAh}\,\frac{\delta}{\cos D}\left[\frac{1}{\dfrac{z_L}{\left(z_L^2-\frac{h^2}{4}\right)^2} \qquad \dfrac{z_u}{\left(z_u^2-\frac{h^2}{4}\right)^2}}\right] e.m.u/cc \qquad (3.2)$$

Where $\frac{1}{S}$ = reciprocal sensitivity (oe/mm deflection at 1.8 meters distance between mirror and scale);

A = cross-sectional area of the specimen (cm^2);

h = height (thickness) of the specimen (cms);

Z_L = distance from the centre of specimen to the centre lower magnet (cms);

Z_u = corresponding distance to the upper magnet;

δ = deflection produced by the azimuthal component of the magnetic vector on the scale at 1.8 meters distance from the mirror (mm);

D = dip of the magnetic vector (degrees).

The above formula holds for the cubes as well and was adopted in the computations of J for the Virgin rock specimens.

Even in the case where the specimen magnetization closely approximates to a point dipole, equation (3.2) does not give an exact value of J, because of the assumption that the magnets are centrally located point dipoles, whereas more than half of the total torque exerted on the lower magnet by the gradient field from the specimen acts on the lower half of its volume. However, even in the case of transverse magnets, where this effect is relatively large, the approximate formula (3.2) can be applied unless the specimen is very close to the lower magnet (in which case other, previously mentioned, errors would also tend to become large). With the existing astatic system (height of the magnet = 0.75 cm) the error in equation (3.2) is about 3% for $Z_L = 3.0$ cm and $< 1\%$ for $Z_L = 6.0$ cms, if Z_L is now taken to be the distance from the centre of the lower magnet to a horizontal point dipole vertically below the magnet, and where the field at the upper magnet has been neglected. In view of these relatively small errors, it was decided to use equation (3.2), though bearing in mind the reduced accuracy obtained when Z_L is small.

3.5 Processing of the Palaeomagnetic measurements

3.5.1 Computation of the mean azimuth and dip

After the azimuth and dip of several specimens from a sample have been obtained, the next step is to compute the mean azimuth and dip for the sample from the specimen data. The process of calculation is straight-forward and can be carried out after the components of the measured vector along the north-south, east-west and vertical directions have been obtained: these are designated X_1, Y_1, and Z_1, respectively. X_1, Y_1, and Z_1 are positive if the north poles of the remanent vector components point north, east and downward respectively. Thus for each speciman:

$$\begin{aligned} X_1 &= \cos A \cos D \\ Y_1 &= \sin A \cos D \qquad (3.3) \\ Z_1 &= \sin D \end{aligned}$$

where A and D are the measured values of azimuth and dip of the remanent vector, X_1, Y_1, and Z_1 are then summed for all specimens of a single sample, and the resulting sums are used to determine the mean vector in the sample.

Thus,

$$\bar{A} = \tan^{-1} \frac{\Sigma Y_1}{\Sigma X_1} \tag{3.4}$$

$$\bar{D} = \tan^{-1} \frac{\Sigma Z_1}{\sqrt{(\Sigma X_1)^2 + (\Sigma Y_1)^2}} \tag{3.5}$$

where $\bar{A}$ and $\bar{D}$ are the mean azimuth and dip of the sample.

Similarly, when the mean direction of magnetization of all samples from a given rock formation at a collection site has been found, the mean azimuth and dip of the formation can be calculated by summing the components of the magnetic vectors in the samples, in an analogous manner to (3.3). Thus,

$$\begin{aligned} X_2 &= \cos \bar{A} \cos \bar{D} \\ Y_2 &= \sin \bar{A} \cos \bar{D} \\ Z_2 &= \sin \bar{D} \end{aligned} \tag{3.6}$$

The resulting values are used to determine the mean vector for the site:

$$\bar{\bar{A}} = \tan^{-1} \frac{\Sigma Y_2}{\Sigma X_2} \tag{3.7}$$

$$\bar{\bar{D}} = \tan^{-1} \frac{\Sigma Z_2}{\sqrt{(\Sigma X_2)^2 + (\Sigma Y_2)^2}} \tag{3.8}$$

where $\bar{\bar{A}}$ and $\bar{\bar{D}}$ now correspond to the mean azimuth and dip of the formation.

If the collections of samples are made from rocks belonging to the same geological formation, but exposed at different sites, the same process can be extended to obtain the mean azimuth and dip of the magnetic vector for the formation.

3.5.2 <u>Statistical analysis of palaeomagnetic data</u>

The data obtained in palaeomagnetic studies consist generally of population of directions of magnetization measured in oriented rock samples from a given geological formation or site. The directions of natural remanent magnetization of rock samples are always more or less scattered for a number of reasons:

(1) The magnetic moments of the ferromagnetic grains in different parts of the rock will not be perfectly aligned. This can be due to a large number of causes: in the first place,

most processes that produce a natural magnetization in rocks probably succeed only in magnetizing a small fraction of the domains present (Irving, Stott, and Ward, 1961) so that the remainder are magnetized randomly, and the total magnetization, in the absence of other errors, consists of vectors randomly scattered about the significantly directed fraction of the domains. Other causes of misalignment are: imperfections in the crystal structure; uneven response to stresses as igneous rocks cool from a high temperature; uneven settling of grains during the formation of sedimentary rocks; chemical action or metamorphism of the rock after it has been formed, etc. With the exception of certain systematic components, such as the inclination error to be expected in some sedimentary rocks (see introduction), which have to be analyzed separately, the naturally occurring misalignment of magnetization directions in a rock formation resulting from many of the above causes, can be lumped together in terms of random scatter.

(2) Changes in the direction of the magnetic field: samples from a rock sequence which represents a geologically long time (say a period or longer) may exhibit differences in their natural remanent magnetization directions, because of the possibility that the palaeomagnetic pole shifted its position during the time span represented by the samples. Hence detailed palaeomagnetic conclusions could not be drawn from a single mean direction of magnetization based upon such a thick sequence. On

the other hand, care must be taken also in averaging the results of a palaeomagnetic study based upon samples representing very short time spans (say, less than a few thousand years), as the secular variation of the geomagnetic field may not have been averaged out in the data in that case (see Section 3.5.3.).

(3) Orientation and measurement errors: these also contribute to the scatter in measured directions (usually random, but possibly including some systematic components, such as human observation errors - see Section 4), and so, provided systematic errors can be considered to be insignificant, the error arising from causes (1), (2), and (3) in the measured directions can be treated as random scatter, subject to the conditions listed below.

Fisher (1953) has developed a method of statistical analysis of palaeomagnetic data in which each magnetic vector is regarded as a point on a sphere of unit radius. Fisher's distribution is similar to a gaussian distribution and is applicable if the population from which the sample is drawn satisfies the two conditions:

(a) The vectors in the population must be distributed with axial symmetry about their mean direction;

(b) The density of the vectors in the population must decrease with increasing angular displacement ψ from the mean direction according to the probability density function:

$$P = \frac{k}{4\pi \sinh k} \exp(k \cos\psi) \qquad (3.9)$$

The quantity k is a constant called the precision parameter and describes the tightness of a group of vectors in the population about their mean direction. High values of k indicate tight groups, while $k = 0$ corresponds to a population uniformily distributed over the entire surface of the unit sphere.

Fisher shows that, provided the above conditions are satisfied, the direction of the vector sum of the N unit vectors of the sample is the best estimate of the true mean direction of the population. The components of the mean direction are:

$$\begin{aligned} X &= \sum_{i=1}^{N} \cos A_i \cos D_i \quad \text{(North component)} \\ Y &= \sum_{i=1}^{N} \sin A_i \cos D_i \quad \text{(East component)} \qquad (3.10) \\ Z &= \sum_{i=1}^{N} \sin D_i \quad \text{(Downward Component)} \end{aligned}$$

and the length of the vector:

$$R = \sqrt{(X^2 + Y^2 + Z^2)} \qquad (3.11)$$

The resultant values of magnetic azimuth, A_R and dip D_R, are:

$$A_R = \tan^{-1}\frac{Y}{X} \quad ; \quad D_R = \sin^{-1}\frac{Z}{R} \qquad (3.12)$$

The best estimate of the precision parameter (for K > 3) is given by:

$$K = \frac{N-1}{N-R} \qquad (3.13)$$

At a probility level of (1-P), the true mean direction of the population lies within a circular cone about the resultant vector R with a semivertical angle $\alpha_{(1-P)}$ given by Fisher (1953).

$$\cos \alpha_{(1-P)} = 1 - \frac{N-R}{R}\left\{\left(\frac{1}{P}\right)^{1/N-1} - 1\right\} \qquad (3.14)$$

It is usual practice in palaeomagnetism to take P as 0.05, which means that there is 1 change in 20 that the true mean direction of the population lies outside the cone of confidence specified by α_{95} and the direction of R.

α_{95} is given approximately by:

$$\alpha_{95} = \frac{140^\circ}{\sqrt{KN}} \qquad (3.15)$$

where the subscript 95 under expresses a percentage.

For all suites of rocks studied in this thesis, the statistical parameters K and have been calculated. For a given N, the scatter in directions decreases for an increase in the value of K (eq. 3.13). For a given scatter within the rock formation, an increase in N, i.e. in the sample from which the mean direction is calculated, results in a smaller cone, and this in turn constitutes a more "reliable" mean direction. However, this is subject to the correct application of the statistical analysis, as already mentioned, and moreover, there are practical upper limits to the number of samples that should be taken from a given rock formation, beyond which no significant improvement in "confidence" can be obtained. This and other aspects of statistical methods in rock magnetism have been discussed by Watson and Irving (1957), Cox and Doell (1960), Cox (1964) and others.

3.5.3 Calculation of the palaeomagnetic pole

Results of the natural remanent magnetization from rock units of the same region can be compared directly. But if results from different regions of the world are to be compared, they must be referred to a common base. One convenient way is to represent data in terms of the geocentric dipole that would produce the measured field direction. This is done by specifying the present geographic co-ordinates of the geomagnetic pole that corresponds to the orientation of this inferred dipole.

In such calculations, the following assumption, which is of fundamental importance in palaeomagnetism, is made: the present geomagnetic field at the surface of the earth may be represented in terms of three components; (1) a relatively small component due to processes occurring above the earth's surface; (2) a component equivalent to the field of a magnetic dipole located at the centre of the earth and inclined $11\frac{1}{2}^{\circ}$ from the axis of rotation; and (3) a non-dipple component, which would remain if the externally produced field and the dipole field were removed. Spherical harmonic analysis of the earth's magnetic field (Chapman and Bartels, 1962) can be applied to resolve these three components separately. The dipole component is by far the largest observed at the surface of the earth, and the one exhibiting the greatest stability with time. Although a dipole inclined at $11\frac{1}{2}^{\circ}$ from the axis of rotation gives the best average fit with the present geomagnetic field, theoretical considerations suggest that the mean position of the dipole component, when averaged over periods of the order of the secular variation (say, a few thousand years) will be along the axis of rotation. This reasoning is supported by palaeomagnetic results from rocks of Pleistocene and Recent age, which indicate quite consistently that the earth's magnetic field, when averaged over a few thousand years, was dipolar in nature and parallel to the present axis of rotation, (e.g., Hospers, 1955). While this experimental test covers only the past million years or so, it provides, along with theory, a plausible basis for extrapolating the "axial geocentric dipole assumption", i.e. it is assumed that in expressing palaeomagnetic results in terms of ancient pole positions, one can represent the ancient earth's magnetic field by an axial geocentric dipole, provided the rocks under study represent a geological sequence spanning at least several thousand years. Moreover, the palaeomagnetic results supporting the axial geocentric dipole assumption also constitute additional evidence for the dynamo theory of the origin of the geomagnetic field (Elsasser, 1956; Bullard, 1949).

Once the mean azimuth and dip of the formation have been calculated, the relative position of the palaeomagnetic pole which would be consistent with the observed mean direction can be obtained by the following equations:

$$\cot p = \frac{1}{2} \tan D \qquad (3.16)$$

$$\sin \lambda' = \sin \lambda \cos p + \cos \lambda \sin p \cos A$$

$$(-90^\circ \leqslant \lambda' \leqslant 90^\circ) \qquad (3.17)$$

$$\varphi' = \varphi + \beta \quad \text{if} \quad \cos p \geqslant \sin \lambda \sin \lambda'$$

$$\varphi' = \varphi + 180 - \beta \quad \text{if} \quad \cos p < \sin \lambda \sin \lambda'$$

$$\sin \beta = \frac{\sin p \sin A}{\cos \lambda'} \qquad (3.18)$$

where p is the angular distance along the great circle from the sampling site to the palaeomagnetic pole;

A is the azimuth (or declination) of the magnetization vector;

D is the dip (or inclination) of the magnetization vector;

λ, φ are the latitude and longitude of the collecting site;

λ', φ' are the latitude and longitude of the palaeomagnetic pole.

From the above relations, a unique palaeomagnetic pole position can be obtained. A more detailed account of the calculation of the palaeomagnetic pole may be obtained from Cox and Doell (1960). It is also customary to express the scatter in the palaeomagnetic results by surrounding the computed pole position with an "oval of confidence", which is a function of the mean dip of the magnetic field at the collection site, and of α_{1-P}. The oval of confidence is expressed in terms of its semiaxes, δp and δm, where δp lies along the great circle passing through the collecting site and the palaeomagnetic pole, and expresses the uncertainty in determining the distance between the latter and the former. δm is perpendicular to the great circle, and corresponds to the uncertainty in azimuth of the computed pole position relative to the collection site.

3.5.4 Correction for the geological tilt

If, subsequent to the acquisition of remanent magnetism, the geological body is structurally deformed, the measured magnetic vector should be transformed to the position in which the beds were originally deposited, which can usually be taken to be the horizontal plane. If the samples were taken from beds that have been tilted, e.g. from the flanks of an anticline, a "tilt" correction" can be made in which the mean magnetization vector, which is considered to be "frozen" within the rocks, is rotated along with the rocks about their strike until the surface of the beds is horizontal. The new direction of the magnetization vector is then assumed to be in the original field direction at the site.

Such a correction is only applicable when there is reason for confidence in the stability of magnetization; i.e. it is assumed that the NRM of the rocks, or the magnetization remaining after "soft" components have been removed in the laboratory, corresponds to the field direction at the time the rocks were formed, and hence prior to geological deformation. Moreover, the tilt correction is only applicable when it can be assumed that there has been no post-depositional rotation of the beds about a vertical axis. This assumption is by no means always justified, and lack of the necessary geological or tectonic history of the relevant rock formations can often set a limit to palaeomagnetic interpretations.

If the above assumptions can be met, the tilt correction (which will be applied to data in Chapters 4 to 6), is calculated as follows:

Let the magnetization before tilt correction (i.e. assuming a flat-lying rock) be split up into three components,

X = north-south (north +)

Y = east-west (east +)

Z = vertical (down +)

The mean components for a number of specimens may be denoted by $\bar{X}$, $\bar{Y}$ and $\bar{Z}$. The resultant mean azimuth and dip are given by:

$$\bar{A} = \tan^{-1} \frac{\bar{Y}}{\bar{X}} \qquad (3.19)$$

$$\bar{D} = \tan^{-1} \left[\frac{\bar{Z}}{\sqrt{\bar{X}^2 + \bar{Y}^2}} \right] \qquad (3.20)$$

The above magnetization direction is split into three components such that,

X_1 is parallel to the strike;

Y_1 is 90° east of strike;

Z_1 is vertical.

The signs are given as follows:

The angle of geological tilt, T, is positive in the direction of downward tilt. Y_1 is taken as positive in that direction and X_1 is thus positive at Y_1-90°. Z_1 is positive downward. The three components can be denoted as:

$$
\begin{aligned}
X_1 &= \cos\theta_1 \cos D \\
Y_1 &= \sin\theta_1 \cos D \qquad (3.21) \\
Z_1 &= \sin D
\end{aligned}
$$

where θ_1 is the angle taken clockwise from X_1, between X_1 and $\bar{A}$.

Let d be the angle which the projection of magnetization on to the Y_1-Z_1 plane makes with the horizontal. Thus d is positive downwards; and

$$d = \tan^{-1}\left(\frac{Z_1}{Y_1}\right) \qquad (3.22)$$

This projection, d, of magnetization is changed due to tilt correction, as follows:

(a) If T is positive, the angle d changes to d+T (since a downward tilt adds to d).

(b) If T is negative, the new angle with the horizontal is d-T (since an upward tilt correction decreases d).

The new components (after tilting) are as follows:

$$
\begin{aligned}
X_2 &= X_1 \\
Y_2 &= Y_1 \frac{\cos(d+T)}{\cos d} \qquad (3.23) \\
Z_2 &= Z_1 \frac{\sin(d+T)}{\sin d}
\end{aligned}
$$

The new value of θ is θ_2, given by:

$$\theta_2 = \tan^{-1}(Y_2/X_2) \qquad (3.24)$$

$$\bar{A}_1 = S + \theta_2 \qquad (3.25)$$

where θ_2 and $\bar{A}_1$ are the new values of θ and $\bar{A}$, respectively, and S is the angle of geological strike.

The new dip, D_1 is given by:

$$\bar{D}_1 = \tan^{-1}\left[\frac{\bar{Z}_2}{\sqrt{(\bar{X}_2)^2 + (\bar{Y}_2)^2}}\right] \qquad (3.26)$$

Thus $\overline{A}_1$ and $\overline{D}_1$ will be the new azimuth and dip of the formation corresponding to the original horizontal of the bedding plane after rotation about strike. Throughout the calculations in this operation, it is easy to make errors of sign, and hence, great care must be taken in order to obtain the correct quadrants for the particular trigonometrical function.

3.6 Errors in palaeomagnetic observations

The observed variations in the palaeomagnetic directions from a suite of rocks may be ascribed to the following sources:

(1) Variation in the orientation of ferromagnetic mineral grains in the rock.

(2) Error in the orientation procedure in the field.

(3) Error in reproducing the field orientations in the laboratory; i.e. misorientations arising when the sample is set in plaster of Paris, during subsequent coring, and in transferring the field orientation marks to the specimen.

(4) Error in the measurements with the astatic magnetometer.

Reference has already been made to contributing causes to the error under (1), and to some extent to those under (2) - (4). The resultant error incorporating all sources will be widely different in different palaeomagnetic investigations, mainly because the error under (1) is a function of so many variables. This resultant error enters into the statistical analysis of the measurements, though such an analysis becomes more meaningful when applied to the "treated" data, i.e. after the original scatter has been reduced by removal of relatively "soft" or viscous components in the laboratory, or by field corrections of the effect of geological deformation. In any case, each palaeomagnetic study requires a separate careful error analysis.

However, the errors under (2) - (4), though contributing to the total scatter in magnetization directions observed for a given suite of rocks, can be separately estimated because orientation and measurement procedures are fairly standardized. Provided these procedures are correctly applied, the scatter in the data contributed by the combined orientation - measurement errors should be similar for different instruments of similar precision. According to Irving (1964) the combined orientation - measurement error is usually in the range 2-5^{o}, which is generally small compared to the observed dispersion between samples from the same collection site. Sometimes the field orientation error above can lie outside this range: e.g. the samples from Virgin Rocks shoals (Chapter 6) could be oriented only to 5-10^{o} accuracy because of the difficulty of orienting the rocks under water, the technique of which has yet to be perfected.

The error under (2) is composed of errors in (i) reading the Brunton compass used to obtain bearings; (ii) determining the horizontal plane; (iii) accurately placing the orientation marks upon the rock surface, which can be quite irregular. The resultant orientation error, largely determined by the error in (iii), is 1-2^{o} in most cases. If the rocks are so strongly magnetic that they significantly deflect a compass in their vicinity, the bearings are obtained by an alternative method (usually compass-sighting at a distance), and this may increase the orientation error by 1^{o} or so. However, because the rocks in this case are strongly magnetic, a relatively smaller measurement error then compensates for the larger orientation error.

The error under (3) can be kept to within 1^{o} by careful work. Then the combined orientation error under (2) and (3) is 2-3^{o} in most cases. All sources of this error contribute more or less randomly to the total observed scatter between samples, but there may be an additional systematic error, (2) (iv), due to inaccuracy in the value of the magnetic declination used to correct magnetic bearings for true north. This value is usually known to $\frac{1}{2}^{o}$ or less, the same azimuth error arising in the case of all samples from the same locality.

Cause (4) includes errors in (i) positioning the specimen in the specimen holder relative to the horizontal plane; (ii) positioning the bearing or azimuth line marked on the specimen perpendicular to the magnetic moment plane of the astatic system; (iii) positioning the vertical axis of the specimen holder (and the vertical axis of the cylindrical receptacle for the specimens) along the vertical axis of the astatic system. The construction of the instrument is such that the specimens are always accurately centered with respect to the vertical axis of the receptacle of the specimen holder; hence no positioning error results from this cause.

Some, but not all, the errors under (4) (i) - (iii) are cancelled when readings taken with the specimen in the upright position are followed by measurements in the upside-down position. Thus, in (i), all azimuth readings will be larger in one position, and smaller in the reversed position than they would have been if the specimen had been correctly centered; the measuring procedure will eliminate most or all of the error with regard to a final azimuth, but both dip and intensity of magnetization measurements will tend to incorporate a net error. Similarly, there will be in general an uncompensated error in azimuth (and a smaller dip error) corresponding to cause (ii). Errors in (iii), due to off-setting from the axis of the astatic system, arise from two causes: (a) the vertical component of the magnetic moment vector will then introduce a non-cancelling, additional horizontal component on the lower magnet, which causes an error in the observed deflection; (b) the effect on the lower magnet, due to a magnetization component it has induced in the specimen, is generally negligible apart from shape effects when the latter is accurately centered underneath the astatic system, but again a significant horizontal field component at the lower magnet can be introduced when the specimen is offset along an axis at right angles to the magnetic moment plane of the astatic system. Errors under (iii) are not compensated by the measuring procedure and can become important; hence great care must be taken to ensure accurate horizontal centering of the specimens below the astatic system. This has been done in the present case.

Apart from a human observation error which may be systematic, the remaining measurement error is that arising from the scale readings, which are individually accurate to 0.25 mm on quiet days. The error from the use of a flat, rather than circular, scale at 1.8 m from the mirror, is also negligible. Eight readings go into azimuth or dip determination, not counting repetitions of the first reading in each of the upright and upside-down positions; these are merely used to distribute the drift of the light spot which generally can be assumed to contribute no significant error. For a specimen producing given deflections on the scale, one can then estimate the error in reading the light spot.

It is more useful, however, to obtain an estimate of the total measurement error. While the various sources are partly random and partly systematic, so that it is difficult to estimate the contribution of each, one can estimate the total measurement error by repeating observations with the same specimen of given intensity of magnetization and noting the scatter in the results.

For this purpose an artificial specimen (pyrrhotite powder set in plaster of Paris) of intensity $J = 2.00 \times 10^{-4}$ e.m.u/cc was subjected to measurements with the astatic magnetometer in the course of several days; before each set of measurements the cylinder was re-positioned in the holder to randomize the error arising from manual positioning. The distance of the specimen centre from the centre of the lower magnet was 6.00 cms. The results were as follows:

Measurement	Azimuth (A)	Dip (D)
1	28.1°	12.0°
2	27.8°	11.8°
3	27.5°	12.1°
4	28.4°	11.3°
5	29.0°	12.3°
6	28.0°	11.6°
7	28.4°	11.2°
Average	28.17°	11.75°

The deviations from the mean are:

for Azimuths -0.1°, -0.4°, -0.7°, $+0.2^\circ$, $+0.8^\circ$, -0.2° and $+0.2^\circ$.

for Dips $+0.2^\circ$, 0.0°, $+0.3^\circ$, -0.5°, $+0.5^\circ$, -0.2° and -0.6°.

The standard deviations of individual measurements are:

$$S_A = 0.48^\circ \qquad S_D = 0.41^\circ$$

Thus it can be said that the direction of magnetization of a specimen having intensity of magnetization of the order of $J = 2 \times 10^{-4}$ e.m.u./cc can be measured to an accuracy in the resultant vector of less than 1° with the astatic magnetometer used by the author. The same reproducibility can also be achieved with weaker specimens (with minimum intensities of the order of, say, 6×10^{-5} e.m.u./cc) measured in closer proximity to the lower magnet; however, other errors, already discussed, may then become prominent.

It should be noted that the actual errors in azimuth and dip, as quoted above for the artificial specimen, are somewhat arbitrary, as the azimuth and dip values themselves are arbitrary; therefore the fact that the individual errors and the standard deviation in the dip are somewhat less than those in azimuth is not significant. (An extreme case would be a specimen with very steep dip; in this case the uncertainty in azimuth could be very large without, however, affecting the standard error in the measurement of the total magnetization vector).

CHAPTER 4

PALAEOMAGNETISM OF THE BASALT FLOWS FROM THE SOUTH COAST OF LABRADOR

4.1 Preliminary remarks

It was stated in the Introduction that the measurement of the magnetic vector in a rock can provide evidence on the direction of the earth's magnetic field at the time of formation of the rock. This knowledge of the geomagnetic field during geological times can be successfully utilized in solving many physical and geological problems. During the summer of 1965, the author carried out a field collection of oriented samples from the basalt formations around Chateau Bay and Table Head, Labrador. A few samples from an associated sedimentary formation were also taken. The results and discussions of the palaeomagnetic investigations of these rocks form the subject matter of this chapter.

4.2 State of knowledge about palaeomagnetic directions of rocks from Newfoundland

Very few palaeomagnetic data relating to Newfoundland rocks have been reported. The results are found in two publications:

(1) Nairn, Frost, and Light (1959) sampled Signal Hill sandstones on the Avalon Peninsula, considered to be of late Precambrian age, as well as the Mississippian Codroy rocks north of Robinson's River on the west coast of Newfoundland. They

measured the palaeomagnetic directions of these rocks and compared the pole positions obtained from their data with the mean pole position of the Precambrian rocks (Hakatai Shales) of the western and central U.S.A. reported by Creer, Irving and Runcorn (1957). In order to bring their results into closer agreement with those from the Hakatai Shales, Nairn et al suggested a counterclockwise rotation of Newfoundland with respect to the rest of North America by 20^{o} in post-Carboniferous times.

Black (1964) collected oriented rock samples from Palaeozoic formations of western Newfoundland, including the Bradore formation (L. Cambrian), Clam Bank group (L. Devonian) and the Codroy group of Mississippian age. He compared the palaeomagnetic directions inferred from these formations with the results from the Ratcliffe Brook formation of New Brunswick (L. Cambrian), the Perry volcanics and sediments of New Brunswick (U. Devonian) and sediments of Prince Edward Island, New Brunswick and Gaspé (Permo-Carboniferous and Carboniferous), respectively. From his results Black concluded that there is no relative rotation between Newfoundland and eastern Canada at least since Carboniferous time. The differences in results (between sites in Newfoundland and eastern Canada) for Cambrian and Devonian rocks were attributed to a counterclockwise rotation of Newfoundland by 30^{o} with respect to eastern Canada during late Devonian age. He suggested that the fulcrum of the 30^{o} rotation of Newfoundland is either at point A or B (Fig. 4), the rotation being anticlockwise in either case.

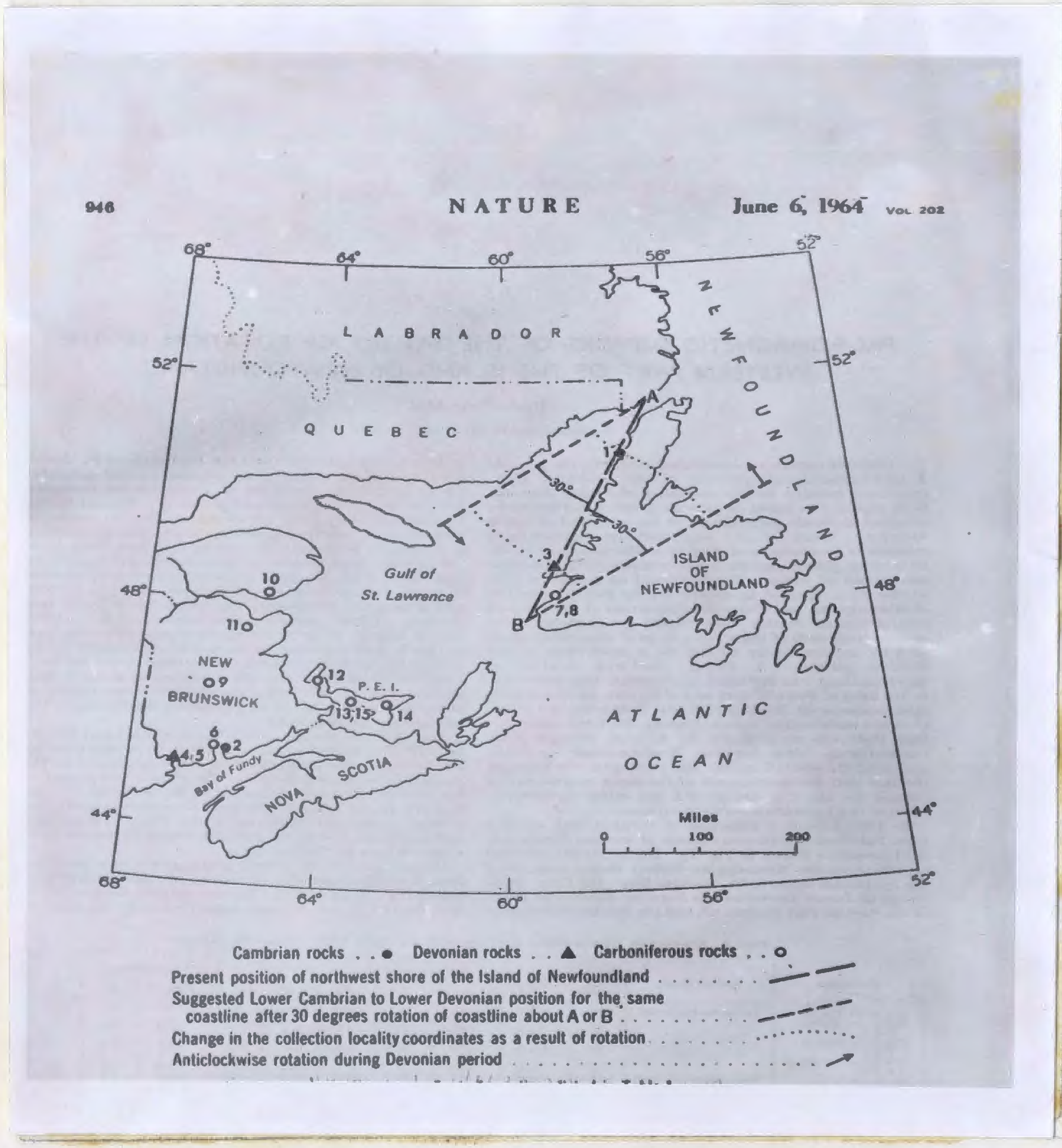


Fig. 4 COUNTERCLOCKWISE ROTATION OF NEWFOUNDLAND DURING LATE DEVONIAN TIMES AS SUGGESTED BY BLACK (reproduced from Black 1964)

However, the interpretation of the data is inconclusive, both in the case of Nairn et al and that of Black:

Nairn et al compared the results from Newfoundland with those from central U.S.A. Irving (1964) comments on their results: "the reasons (for suggesting the counterclockwise rotation by 20^o) are not compelling since the time correlation with the Hakatai Shales is conjectural, and the discrepancy could, in any case, arise from the limited sampling of these beds".

Black compared the pole positions from western Newfoundland with those from Prince Edward Island and New Brunswick. However, these formations are far away from the fulcrum of the suggested rotation, and moreover, there is some possibility (e.g. Prof. Marshall Kay of Columbia University - private communication) that part of the Atlantic land region of Canada might have rotated relative to the remainder of the North American mainland. If this were the case, the rock suites from Prince Edward Island, Gaspé and New Brunswick might be ill suited for testing the hypothesis of the rotation of Newfoundland relative to the North American mainland as a whole.

4.3 Objectives of the present study

The basalt flows from the south coast of Labrador were chosen for the present investigations. There are ~~three~~ two main reasons for this choice; apart from the possibility that the study might provide further evidence on the rotation of Newfoundland:

(1) Few reliable palaeomagnetic results based on lower Palaeozoic rocks have been reported from mainland North America and they include only three Cambrian results (Howell, Martinez and Statham, 1958; Collinson and Runcorn, 1960; and Black, 1964). On the other hand, Cambrian rocks are relatively well represented in Newfoundland, and it would be useful to add further Cambrian palaeomagnetic evidence to the single result already published, based on the Bradore formation (Black 1964). Even if the Labrador rocks selected for the present investigation turn out to belong to a period other than the Cambrian (see below) it is of interest to study any palaeomagnetically promising rock formation in Newfoundland that can add to the still sparse North American evidence relating to other periods of the Palaeozoic era, and possibly also more recent times. Among the most promising raw material for palaeomagnetism are relatively unmetamorphosed volcanic rocks which have not undergone substantial geological deformation and are, preferably, associated with a sedimentary series also of palaeomagnetically suitable rock. At first sight the rocks overlying the Precambrian metamorphics at Henley Harbour and Table Head offer all these conditions which occur relatively rarely in combination, though it appears now that both the basalts and the red sediments underlying them are more highly metamorphosed than was evident before they were more closely examined (see Section 4.12).

(2) Instead of the usual procedure in palaeomagnetism,

which is to determine the ancient directions of the geomagnetic field from measurements in independently dated rocks, palaeomagnetic results are sometimes used for the converse purpose of drawing conclusions regarding the age of the rocks. Such inferences can rarely be very accurate, as they must be based on previously determined pole positions relative to the land mass containing the collection site, and these are still only crudely known, particularly in the case of the lower Palaeozoic rocks in North America. On the other hand, the three Cambrian pole positions relative to mainland North America and the single pole relative to Newfoundland are all in low latitudes in the eastern hemisphere, which is drastically different from, say, a Tertiary pole which would tend to be in a high latitude. Hence, magnetically reliable result from the Labrador rocks could certainly be used to draw a significant distinction between a Cambrian and a Tertiary age, though it would be doubtful whether a distinction between the Cambrian and the middle Palaeozoic (say, the Devonian) could be drawn; this difficulty is increased because the basalt exposed in Henley Harbour and Table Head probably represents only a single flow, so that the secular variation of the ambient earth's field is unlikely to have been averaged out in the magnetization of these rocks. Even if the latter were perfectly stable it would probably deviate from the mean field direction at that time, averaged over some thousands of years.

Nevertheless, the age of these rocks is sufficiently uncertain to make it interesting to attempt such a test: a number of geologists have visited the area around Henley Harbour and Table Head, 14km to the north. Kranck (1939) reported that the basalt sheets of the Chateau Bay (Henley Harbour) area are of Palaeozoic age. Christie (1951) and Douglas (1953) believe that both the basalts and the underlying sediments are of Palaeozoic age and Eade (1962) mapped them accordingly. Christie inferred that these rocks probably belong to the lower Cambrian, because of the close association of basalts and sediments and the fact that the latter resemble the arkose and conglomerate of the Bradore formation exposed in its type location at Forteau Bay, further south along the Labrador coast and placed in the lower Cambrian because of fossil evidence (Schuchert and Dunbar, 1934). Neither the basalt-sediment association, nor the resemblance with the dated Bradore sediments is conclusive, however. In particular, no fossils of any kind have been found at Henley Harbour or Table Head and further argillaceous and calcareous members of the Bradore formation exposed further south are absent at these localities.

Wanless et al (1965) conducted some Potassium-argon age determinations of the Table Head basalts and obtained a middle Devonian age. While this result is also not necessarily conclusive, a post-Cambrian age for the Henley Harbour and Table Head rocks cannot be ruled out.

4.4 Geology of the area

In the Chateau Bay area, the basalt formations are massively exposed at three localities: Devil's Dining Table (Henley Island), Castle Island, and Table Head. Henley Island and Castle Island are both at Henley Harbour and are separated by a tickle, while Table Head is about 15km north-east of Henley Harbour. Fig. 5 shows the three localities.

The exposure named "Devil's Dining Table" is a 24-metre section of fine-grained black basalt showing good columnar jointing. The basalts rest unconformably on Precambrian gneisses, except in one locality where Kranck (1939) noted an underlying lens of reddish brown arkose. Castle Island has a basalt exposure of similar thickness which obviously is part of the same sheet. In thin section the basalts contain augite and labradorite with minor alteration products set in a fine-grained opaque ground mass. The flow at both localities was sharply divided by a break about half-way (12m) from the base, but this showed no evidence (e.g. vesicular structure) of constituting the division between 2 flows. It was therefore believed to be more likely that the basalt constituted a single, 2-tier flow. Strike and dip were obtained by noting the trend of the dividing plane, but in other places (and preferably) by measuring the direction and angle of inclination of the basalt columns. Strike and dip measurements were made at various places and it was found that, on the average, the basalts have a strike of 20^{o} and a dip of 14^{o}SSE; with an estimated uncertainty of about $\pm 10^{o}$ in the strike value and about $\pm 3^{o}$ in the dip value.

Table Head, as its name suggests, is a flat hill, where

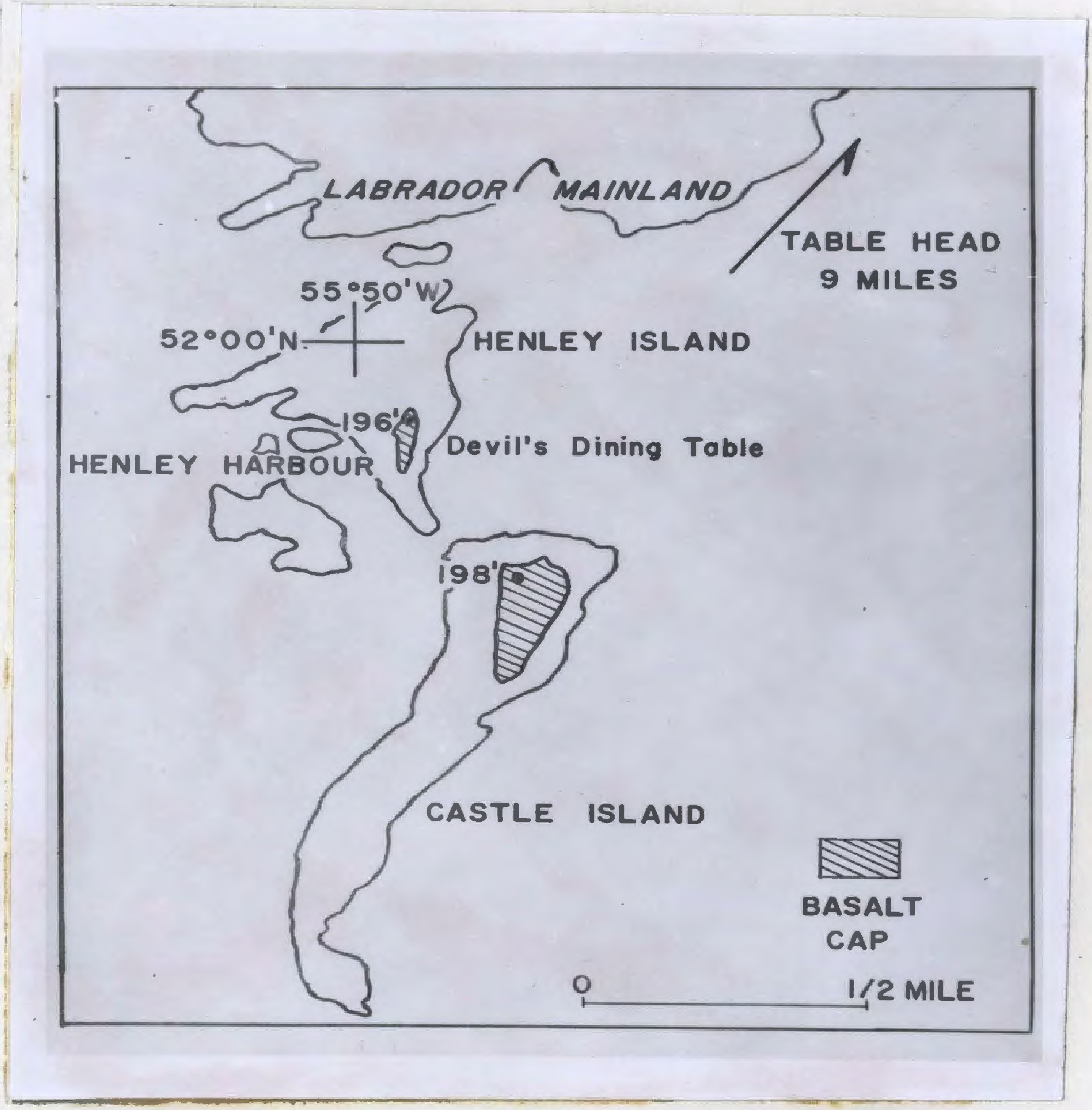


Fig. 5 LOCALITIES SAMPLED FROM THE SOUTH COAST OF LABRADOR

18 meters of reddish arkose and conglomerate are surmounted by a basalt flow 3.6 meters thick. The basalt top is flat as at Henley Harbour, but columnar structure seems to be completely absent. Only a single flow, which could be an extension of the flow at Henley Harbour, can be distinguished. The general trend of the basalt hill is NE-SW. The position of the basalts at Table Head appeared to be almost horizontal, but exact values of strike and dip were not measured during the 1965 field collection; however, ~~Such~~ measurements were made in some detail by members of the Physics Department during a subsequent summer field trip, in 1966.

4.5 Field techniques

A total number of 36 oriented samples were collected from the three areas of Henley Harbour, Table Head, and Castle Island, but only 32 of these were prepared for measurement. The sampling was conducted in such a manner that the average distance between the samples was 25 meters. In addition, these samples were collected from different vertical levels within the formation. Only five samples of sedimentary rock (3 from Henley Island and 2 from Table Head) are included in the collection.

The following table gives the collection particulars:

			No. of Samples Collected		
Site	Latitude	Longitude	Basalts	Sediments	Total
Henley Island	$51^{\circ}59\frac{1}{2}'$N	$55^{\circ}51\frac{1}{2}'$W	20	3	23
Castle Island	$51^{\circ}59'$N	$55^{\circ}51\frac{1}{2}'$W	6	0	6
Table Head	$52^{\circ}15'$N	$55^{\circ}40'$W	5	2	7

The samples were oriented by the standard technique, using a Brunton compass and a level. A minimum of two horizontals were marked on each sample. Usually one of these, but in some cases an additional horizontal on the top, served as the bearing line.

In the laboratory, each sample was set in plaster of Paris in a preferred position related to its field orientation. Cores were drilled with a drill press using a water-cooled diamond drill bit of 2.2cm internal diameter. The cores were then trimmed with a diamond saw to obtain short cylinders 2.2cm in diameter and 2.2cm in length. The orientation error due to setting in plaster, and during transfer of the orientation marks from sample to core and to the specimen surfaces, was probably 1° relative to the horizontal plane. The fact that results from specimens cut from two to three cores, rather than a single core were usually averaged to give the mean direction of magnetization for sample, tends to compensate for the error in transferring marks (but not for the other orientation errors).

4.6 Measurement of the magnetization directions

From the 32 samples a total number of 193 specimens were obtained. The direction of the magnetic vector and the intensity of magnetization, J, were measured for each specimen by the procedure outlined in Section 3.3. The computed values of azimuth A, dip D, and intensity J for the 193 specimens are presented in Table 6.

TABLE 6

REMANENT MAGNETIZATION OF THE BASALT SPECIMENS

FROM HENLEY HARBOUR AND TABLE HEAD, LABRADOR

SPECIMEN NO.	AZIMUTH A (degrees)	DIP D (degrees)	INTENSITY J (e.m.u./cc x 10^{-4})
HH1 A1	87	44	5.2
A2	77	44	9.1
A3	77	65	10.1
B2	99	43	6.7
B3	82	42	6.4
C1	109	27	6.2
C2	113	28	4.5
C3	131	21	5.7
HH3 A1	167	75	3.3
B1	54	73	4.1
B2	350	66	5.5
C1	54	74	4.9
C2	138	69	3.9
D1	49	71	1.3
D2	3	60	5.0
D3	350	62	3.9
HH4 A1	331	74	6.1
A2	301	73	5.1
A3	348	72	3.5
B	336	62	5.4
C1	309	68	8.2
C2	324	69	4.7
C3	283	60	3.9
D1	315	53	5.0
HH6 A1	100	74	7.3
A2	118	78	4.3
B1	96	60	3.9
B2	100	63	4.5
HH7 A1	285	59	4.8
A2	277	78	4.6
A3	284	66	2.4
A4	291	60	3.6
HH8 A1	218	-65	5.6
A2	222	-67	5.5
B1	76	72	1.1
B2	84	63	3.3
C1	205	-57	3.0
C2	241	-31	4.9
C3	250	-45	4.1
D2	227	-62	1.9

TABLE 6 (CONTINUED)

SPECIMEN NO.	AZIMUTH A (degrees)	DIP D (degrees)	INTENSITY J (e.m.u./cc x 10^{-4})
HH9 A1	315	- 4	472
C1	330	- 1	487
HH 10 A1	141	83	8.6
A2	101	79	3.6
B1	10	78	5.1
B2	144	82	5.3
C1	41	69	5.7
C2	45	69	4.5
C3	358	71	2.2
D1	353	80	3.0
D2	319	79	4.5
HH 11 A1	20	47	4.2
A2	18	68	6.1
A3	10	77	4.9
B2	5	62	5.3
B3	14	68	4.9
C1	359	51	6.2
C2	5	63	5.0
D1	20	60	8.2
D2	21	58	3.0
HH 12 A1	31	51	2.8
B1	5	44	3.8
B2	358	40	2.5
D1	28	66	2.0
D2	15	57	3.8
E1	2	56	5.6
E2	8	41	3.6
F1	230	74	3.5
F2	355	64	1.3
G1	20	69	4.6
G2	352	52	4.4
H1	358	73	1.9
H2	18	54	3.1
I1	336	35	2.0
I2	4	47	2.9
HH 13 A1	288	77	8.5
A2	281	80	7.6
A4	320	83	6.8
A5	285	78	5.5
B1	295	72	7.1
B2	292	69	6.5
B3	322	70	6.5
B4	327	72	5.9

TABLE 6 (CONTINUED)

SPECIMEN NO.		AZIMUTH A (degrees)	DIP D (degrees)	INTENSITY J (e.m.u./cc x 10^{-4})
HH 14	A1	341	69	4.7
	A2	0	71	7.2
	B1	350	77	6.7
	B2	47	75	6.5
	C1	351	73	5.9
	C2	320	79	7.2
	D	346	78	3.2
	E	341	75	5.0
HH 16	A1	78	77	10.1
	A2	26	83	12.3
	B1	347	60	5.6
	C1	93	76	5.7
	C2	87	75	8.7
	D1	98	74	14.0
	D2	103	75	4.6
HH 17	B1	15	64	6.8
	B1	2	66	5.2
	C1	2	65	5.1
	C2	348	51	4.7
	D1	354	41	4.1
	D2	357	44	5.2
HH 18	A1	342	62	7.5
	B1	355	75	16.3
	B2	23	84	6.2
	C1	36	86	17.1
	C2	17	84	9.7
	D1	349	69	13.4
HH 19	A1	36	56	5.0
	A2	90	66	5.3
	B1	17	66	4.0
	B2	349	54	4.2
	C1	48	69	4.7
	D1	4	67	3.6
	E1	339	70	5.7
HH 20	A1	25	72	6.7
	A2	24	73	7.2
	A3	32	75	14.7
	A4	27	75	13.4

TABLE 6 (CONTINUED)

SPECIMEN NO.		AZIMUTH A (degrees)	DIP D (degrees)	INTENSITY J (e.m.u./cc x 10^{-4})
HH 21	A1	155	79	1.9
	A2	314	72	3.5
	A4	198	73	2.3
	B1	334	66	3.1
	B3	172	72	4.0
HH 22	A1	338	73	17.2
	A2	344	75	7.9
	B1	0	73	10.6
	B2	6	72	9.9
	B2	26	73	10.8
	B4	29	79	12.5
HH 23	A1	39	69	13.8
	A2	6	73	12.5
HH 24	A1	41	69	12.0
	B1	75	69	9.0
	B3	45	73	10.3
	C1	43	72	11.1
	C2	32	75	12.3
	D1	51	69	11.6
HH 25	A1	18	76	4.3
	A2	16	75	3.8
	A3	331	77	3.0
	B1	5	76	6.8
	B2	33	66	5.2
	B3	55	65	4.3
HH 26	A1	337	68	3.4
	A2	307	58	9.4
	A3	314	60	6.0
	B1	325	57	9.2
	B2	325	61	5.8
	C2	332	59	8.8
	C3	323	60	9.7
TH 2	A1	64	60	13.0
	A2	89	74	13.2
	B1	54	57	10.0
	B2	70	49	15.4
	B3	86	40	12.0
	B4	57	47	9.2
	C1	64	35	17.4
	C2	63	50	23.0
	C3	77	43	17.9
	D1	77	68	10.6
	D2	34	74	22.5

TABLE 6 (CONTINUED)

SPECIMEN NO.	AZIMUTH A (degrees)	DIP D (degrees)	INTENSITY J (e.m.u./cc x 10^{-4})
TH 4 A1	216	57	20.1
A2	216	54	38.8
B1	228	17	194
B2	270	27	153
B3	269	48	129
C1	255	68	29.2
C2	263	50	37.6
C3	262	56	41.6
TH 5 A1	78	55	11.3
A4	49	71	13.0
C1	80	61	14.8
C2	54	57	17.9
C3	45	44	13.9
C4	40	39	18.4
C5	38	63	25.2
D2	57	71	9.3
D3	66	76	15.9
TH 6 A2	118	86	9.6
A3	60	81	12.1
B1	55	33	3.5
B2	97	52	6.9
B3	102	43	14.6
C1	98	36	3.3
C2	89	28	13.2

TABLE 6A

REMANENT MAGNETIZATION OF THE SEDIMENTARY SPECIMENS

FROM HENLEY HARBOUR AND TABLE HEAD, LABRADOR

HH 27 A1	32	63	0.81
A2	57	63	1.42
HH 28 A2	44	81	1.01
A3	19	84	1.13
TH 7 B1	303	76	0.33
C1	323	70	1.37
E1	300	72	0.99

4.7 Analysis of the specimen data

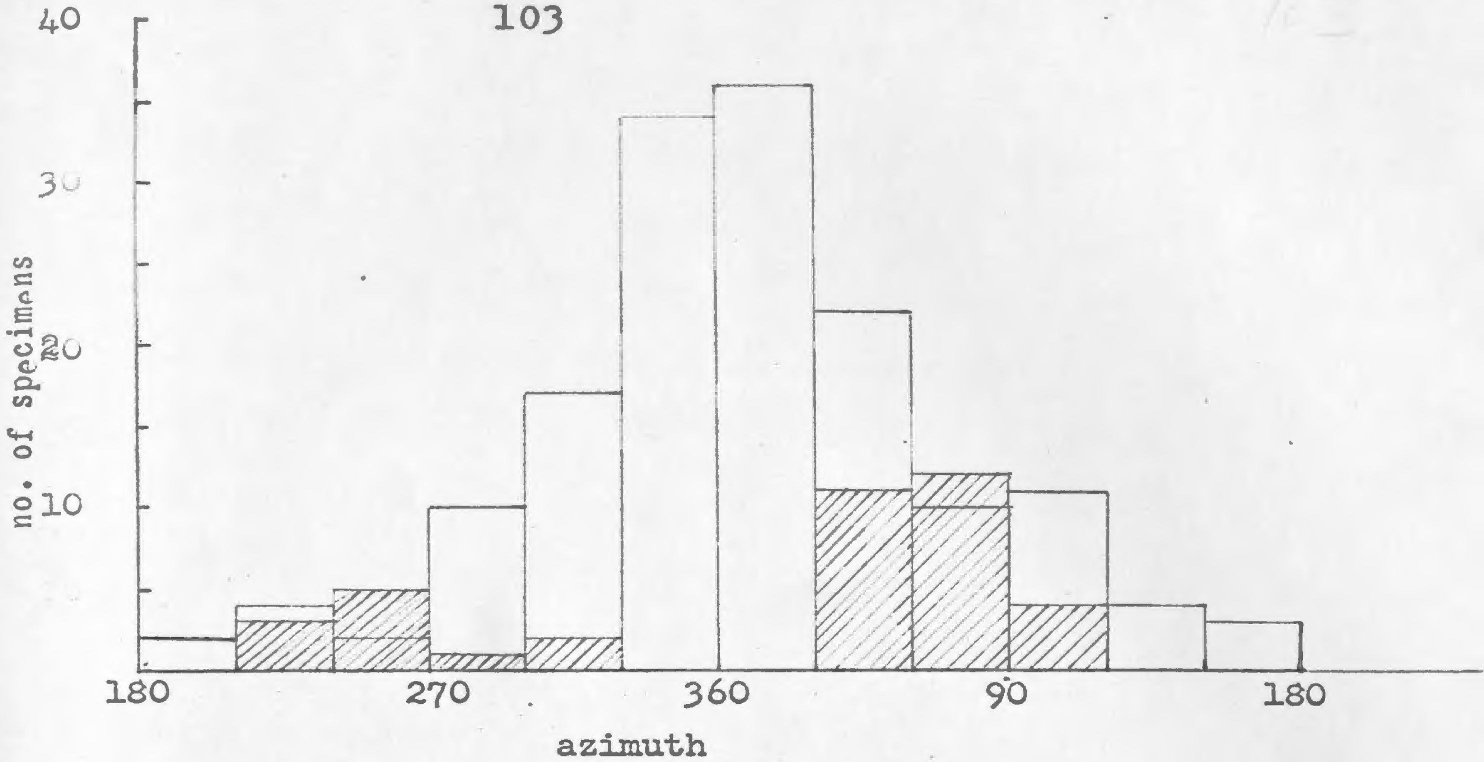
4.7.1 Remanent magnetic azimuth and dip

The azimuths A of the NRM, obtained for the 193 specimens show good consistency within the limits of experimental errors. There are however a few noticeable deviations. A histogram of azimuth values (Fig. 6.a) has its peak in the range 0 to 30°. About 55% of the specimens have their azimuths between 330° and 60°.

The dips of the NRM show greater consistency than the azimuths, which is to be expected since most of the dips were greater than 60°. All the dips are downward except six out of a total of 193 specimens: four of these are from Sample HH8 and two from HH9. As the intensity of magnetization of HH9 was unusually large ($J = 4.8 \times 10^{-2}$ emu/cc), and as only 4 of the 6 specimens from HH8 show this anomalous behaviour, it probably carries little significance. A histogram of dips is given in Figure 6.b. One third of the dip values (67) are in the range 70° to 80°. It may be seen from Fig. 6.b and Table 6, that the dips of the magnetic vector in Table Head basalts tend to be somewhat lower than those in the Henley Harbour rocks.

4.7.2 Remanent intensity of magnetization

Values of the remanent intensity J for the basalts, listed in Tables 6, 7 and 8, encompass a wide range (1.3×10^{-4} to 2.5×10^{-3} emu/cc, excluding HH8 and HH9). This is not unusual



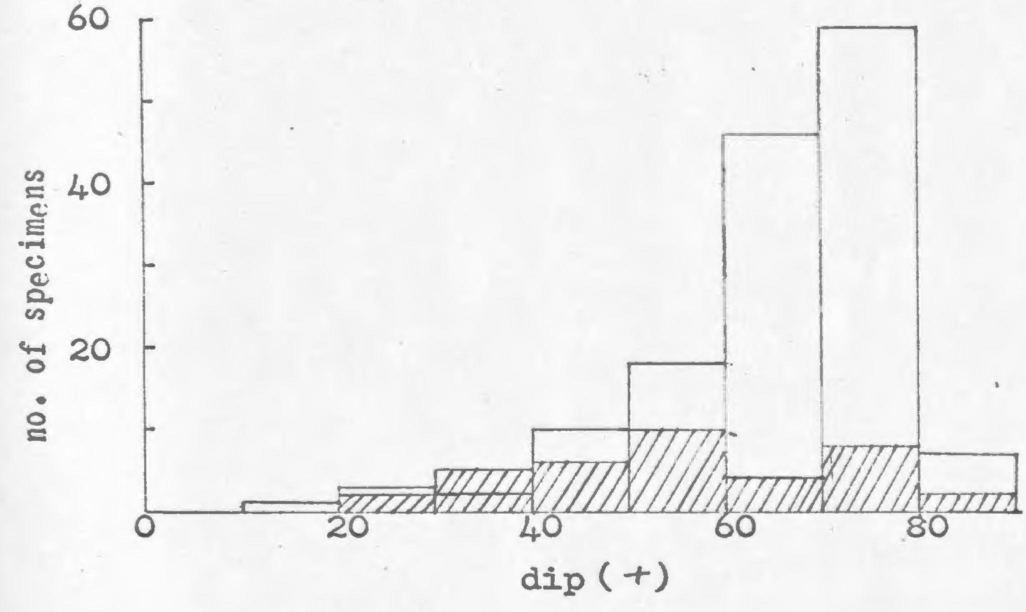


Fig. 6 HISTOGRAMS OF REMANENT MAGNETIC (A) AZIMUTH AND (B) DIP OF BASALTS FROM HENLEY HARBOUR AND TABLE HEAD

☐ HENLEY HARBOUR ▨ TABLE HEAD

even for samples from a single volcanic flow, but the variation between specimens from the same sample also tends to be quite large and, as might be expected, a larger than usual scatter in J values for the same sample tends to be associated with a larger scatter in directions. This is seen in Table 6 (e.g. Samples HH3, HH10, HH12, and TH6) and Table 7. (It should be noted that the precision parameters K and α_{95} are here computed only as a measure of confidence in the mean direction at a <u>site</u>: such estimates are not too meaningful when applied to a very small number of directions, as for example the two specimen values averaged in the case of sample HH23; hence the statistical data in Table 7 should be considered with this limitation in mind). Even sample HH12, where 15 specimens were obtained from locations only a few cms apart within the rock, yielded a fairly wide scatter in directions, and intensities varying by more than a factor of 4. This relative inhomogeneity in the magnetization may be the reflection of partial chemical alteration in the rock.

In general, the Table Head basalts are more magnetic than the Henley Harbour basalts. Two separate histograms are plotted for the two localities and are shown in Figure 7. The majority of the Henley Harbour basalts have their intensities between $J = 4 \times 10^{-4}$ and 6×10^{-4} emu/cc; on the other hand, for 1 in 4 of the Table Head basalts J is between 1.2×10^{-3} and 1.4×10^{-3} emu/cc.

The sedimentary samples from both areas have a mean intensity of about 9×10^{-5} emu/cc. A fairly large variation in

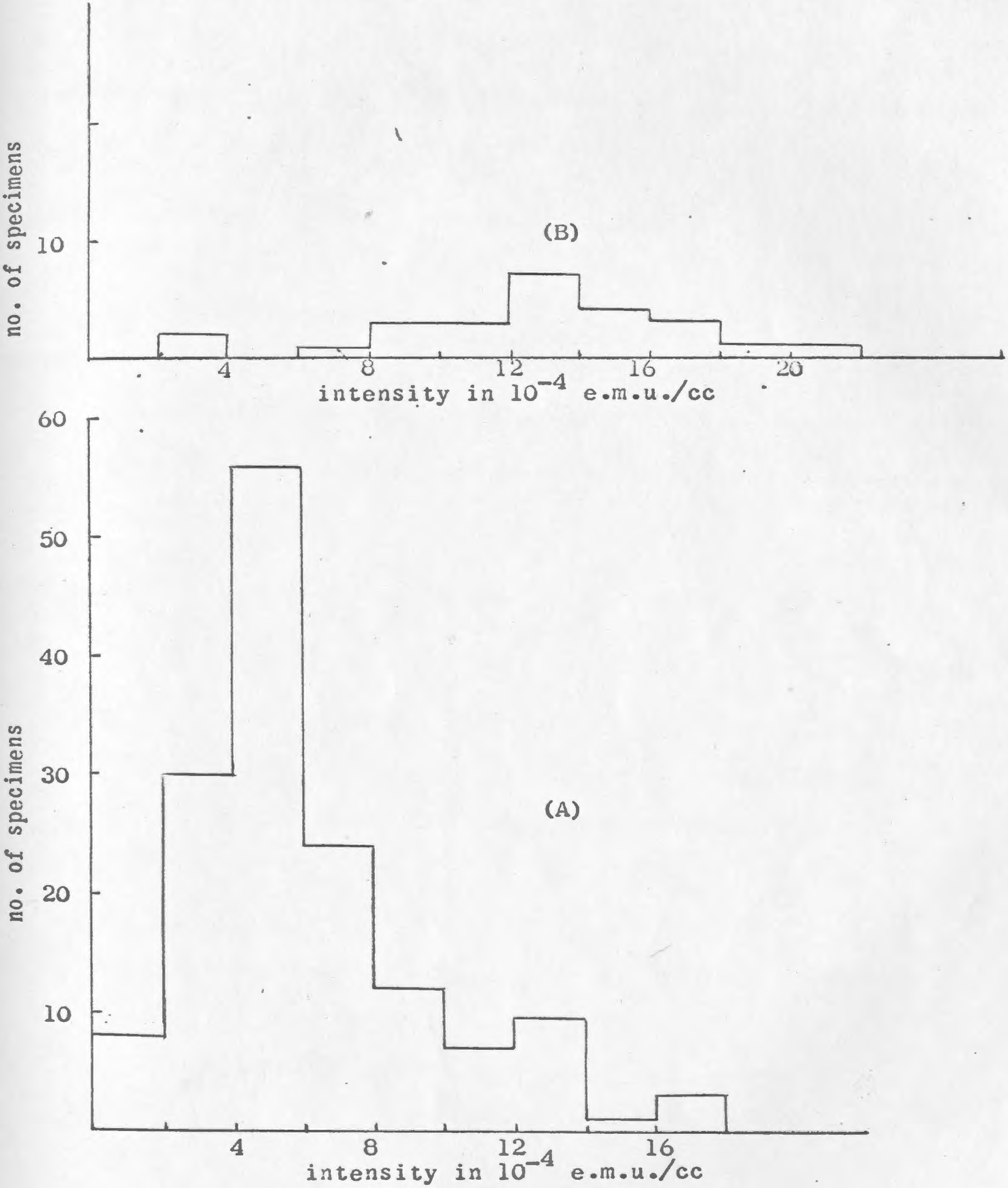


Fig. 7 REMANENT MAGNETIC INTENSITIES OF BASALTS FROM (A) HENLEY HARBOUR AND (B) TABLE HEAD

(excluding specimens TH 4 B1-3)

intensity among specimens from the same sample is also observed in the case of sedimentary rocks. This may be due to some small basalt pebbles enclosed in the sedimentary rocks in some places, particularly in sample TH3 which was not measured for this reason.

As already noted, sample HH9 has an anomalous magnetization. Although the hand sample is not significantly different in appearance from the others, its anomalous behaviour is perhaps attributable to the acquisition of isothermal remanent magnetism by lightning. It is not possible to find a conclusive explanation for the behaviour of HH8 until more exhaustive demagnetization tests have been made.

4.8 Computing the mean azimuth and dip for samples from specimen data

It has been shown in Section 3.5.1 how the mean values of azimuth and dip can be calculated from the specimen data. The mean directions for the samples are tabulated in Table 7. A computer programme set up with the help of G. W. Pearce of the Physics Department has been used for this purpose. A convenient graphical method to present the directions of magnetization involves the use of Schmidt's equal area net and this has been used to plot the mean directions of magnetization of all samples (Fig. 8).

4.9 Computing the mean azimuth and dip for sites out of sample data

The mean azimuth and dip for the basalts of the three sites were computed by the method outlined in Section 3.5.1, and

TABLE 7

REMANENT MAGNETIZATION OF THE BASALT SAMPLES

FROM HENLEY HARBOUR AND TABLE HEAD, LABRADOR

SAMPLE NO.	NO. OF SPECIMENS	AZIMUTH (deg.) $\bar{A}$	DIP (deg.) $\bar{D}$	INTENSITY (e.m.u./cc x 10^{-4}) $\bar{J}$	ESTIMATE OF PRECISION PARAMETER K	RADIUS OF "CIRCLE OF CONFIDENCE" α 95 (deg.)
HH 1	8	99.3	40.6	6.74	16	12.5
HH 3	8	33.2	77.2	3.99	17	11.9
HH 4	8	316.6	66.9	5.24	56	6.6
HH 6	4	100.2	68.7	5.00	82	7.7
HH 7	4	284.5	65.1	3.85	81	7.8
(HH 8)	(8)	(221.4)	(-48.0)	(3.67)		
(HH 9)	(2)	(322.0)	(- 2.0)	(480)		
HH 10	9	31.9	81.0	4.72	41	7.3
HH 11	9	11.7	60.9	5.31	64	5.8
HH 12	15	3.1	56.8	3.16	19	8.4
HH 13	8	301.3	75.1	6.80	137	4.2
HH 14	8	351.6	75.2	5.80	124	4.4
HH 16	7	64.8	78.5	8.71	32	9.4
HH 17	6	358.0	54.8	5.18	43	8.7
HH 18	6	353.8	77.0	11.70	57	7.6
HH 19	7	20.5	67.8	4.64	24	10.8
HH 20	4	26.5	73.5	10.50	1590	1.8
HH 21	5	244.3	84.9	2.96	17	15.3
HH 22	6	1.9	74.6	11.48	171	4.4
HH 23	2	41.7	70.6	13.15	815	3.5

TABLE 7 (CONTINUED)

SAMPLE NO.	NO. OF SPECIMENS	AZIMUTH (deg.) $\bar{A}$	DIP (deg.) $\bar{D}$	INTENSITY (e.m.u./cc x 10^{-4}) $\bar{J}$	ESTIMATE OF PRECISION PARAMETER K	RADIUS OF "CIRCLE OF CONFIDENCE" α_{95} (deg.)
HH 24	6	47.9	71.5	11.05	240	3.7
HH 25	6	21.9	73.9	4.56	65	7.1
HH 26	7	322.5	60.3	7.47	176	4.0
TH 2	11	67.3	54.2	14.92	27	8.1
TH 4	8	247.3	49.0	32.12	12	14.1
TH 5	9	54.6	60.2	15.52	30	8.5
TH 6	7	86.5	52.0	9.02	10	17.1

TABLE 7A

REMANENT MAGNETIZATION OF THE SEDIMENTARY SAMPLES

FROM HENLEY HARBOUR AND TABLE HEAD, LABRADOR

HH 27	2	43.8	63.1	1.12	92	[illegible]
HH 28	2	31.3	82.5	1.07	25	[illegible]
TH 7	3	308.8	72.6	0.89	243	[illegible]

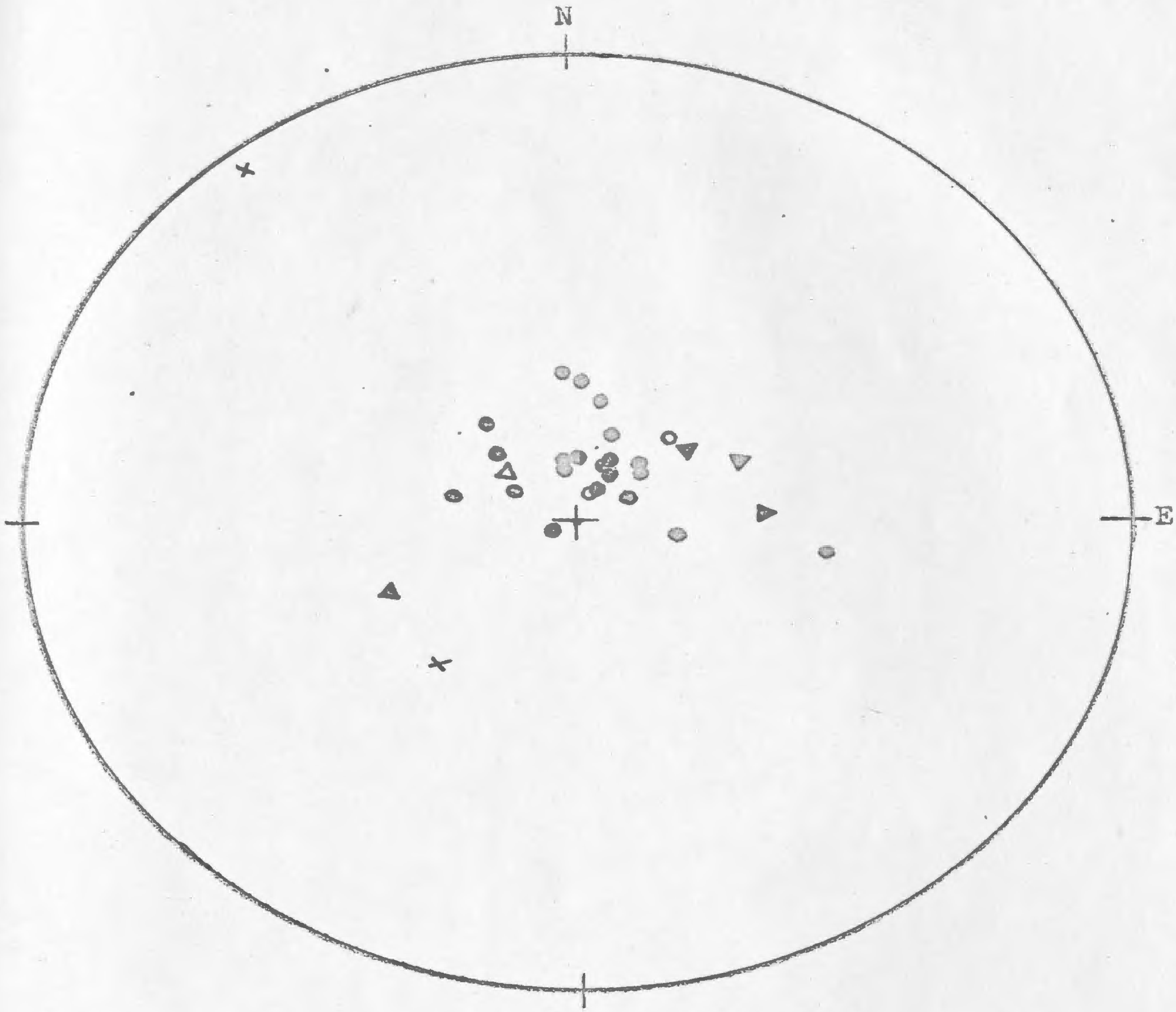


Fig. 8 EQUAL AREA NET OF MAGNETIZATION OF ROCK SAMPLES FROM LABRADOR

- ● Henley Harbour basalts (Dips downward)
- O Henley Harbour sediments (" ")
- ▼ Table Head basalts (" ")
- ▽ Table Head sediments (" ")
- ✗ Samples HH 8 and HH 9 with dips upward

TABLE 8

MEAN MAGNETIC VECTORS OF THE BASALTS FOR THE THREE SITES

SITE	MEAN AZIMUTH (deg.) $\bar{A}$	MEAN DIP (deg.) $\bar{D}$	MEAN INTENSITY (e.m.u./cc x 10^{-4}) $\bar{J}$	PRECISION PARAMETER K	RADIUS OF "CIRCLE OF CONFIDENCE α_{95}	NO. OF SAMPLES
Henley Harbour	13.1	76.1	6.55	16	9.0	15
Castle Island	9.3	73.9	7.45	42	8.8	6
Table Head	72.6	72.8	17.89	5	31.9	4

MEAN MAGNETIC VECTOR FOR THE BASALT FORMATIONS OF LABRADOR

21.2	76.5	8.54	15	7.6	25

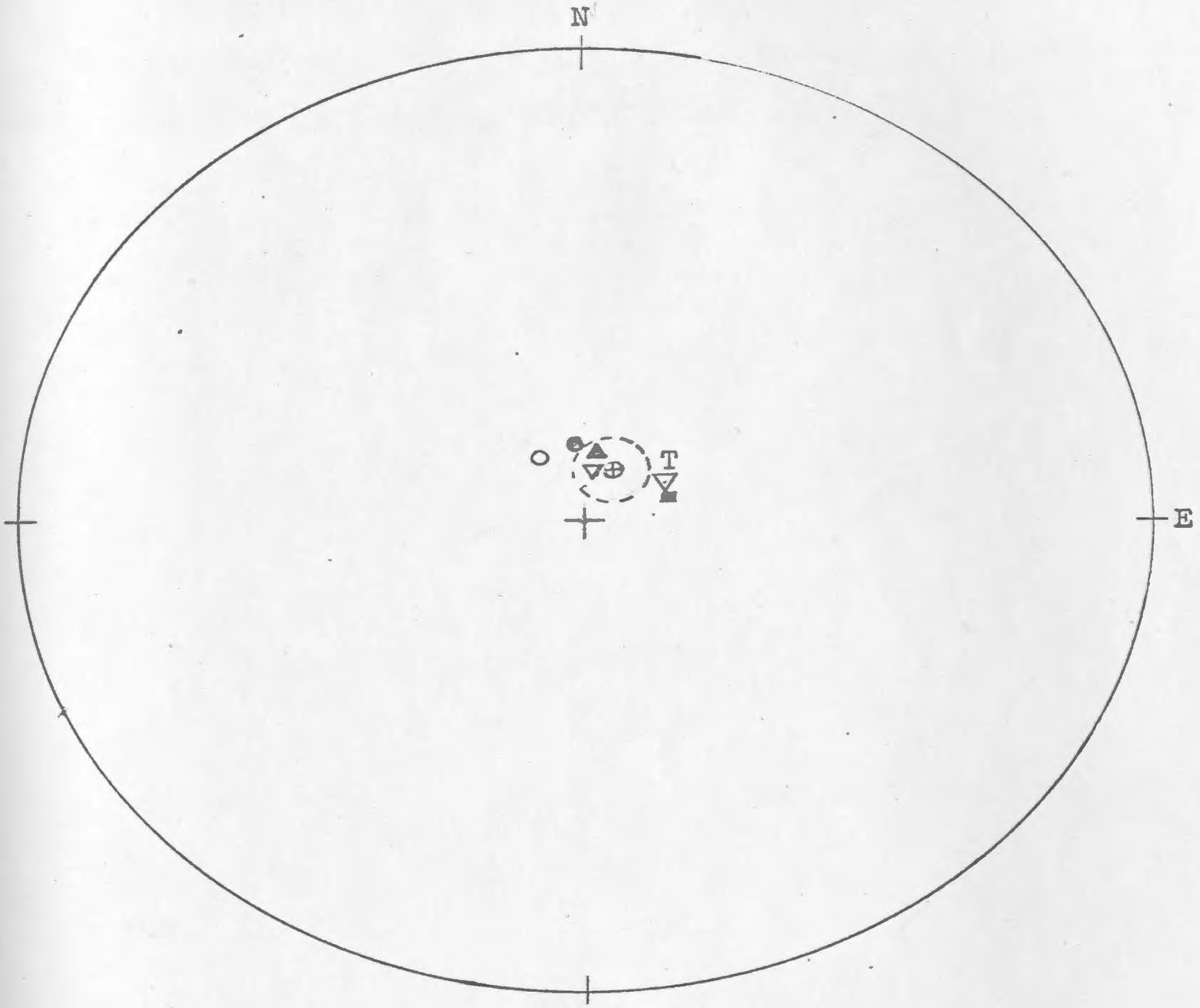


Fig. 9 MEAN DIRECTIONS OF MAGNETIZATION OF BASALT FLOWS FROM LABRADOR

▽ Basalts from Henley Harbour (North poles downward)
▲ Basalts from Castle Island (" " ")
■ Basalts from Table Head (" " ")
⊕ Mean for the three sites
● Theoretical dipole field at sampling site
○ Direction of the present geomagnetic field at sampling site
▽^{T} Henley Harbour results after tilt correction

95 % Circle of confidence

the results are given in Table 8. In these calculations, each sample mean vector is given unit weight. Statistical parameters K and α_{95} were calculated for each site according to equations 3.13 and 3.15 and are tabulated in columns 5 and 6. The mean vectors for the three sites are plotted in Figure 9.

4.10 Computing the mean azimuth and dip for the basalt formation from sample data

From the mean site data in Table 8 the mean magnetic vector of the basalt formation was computed according to Section 3.5.1. The calculation for the three sites combined, and with unit weight assigned to each sample, is given below.

Sites	No. of basalt samples	ΣX	ΣY	ΣZ
DDT, CI, TH	25	5.06	1.96	22.58

$(\Sigma X)^2 = 25.6 \qquad (\Sigma Y)^2 = 3.9 \qquad (\Sigma Z)^2 = 509$

$(\Sigma X)^2 + (\Sigma Y)^2 = 29.5 \qquad \sqrt{(\Sigma X)^2 + (\Sigma Y)^2} = 5.43$

$$\text{Mean azimuth} = \tan^{-1} \frac{\Sigma Y}{\Sigma X} = 21.2^\circ$$

$$\text{Mean dip} = \tan^{-1} \frac{\Sigma Z}{\sqrt{(\Sigma X)^2 + (\Sigma Y)^2}} = 76.5^\circ$$

$$R = \sqrt{(\Sigma X)^2 + (\Sigma Y)^2 + (\Sigma Z)^2} = 23.4$$

$$K = \frac{N-1}{N-R} = 15.0$$

$$\alpha_{95} = \frac{140}{\sqrt{NK}} = 7.6^{\circ}$$

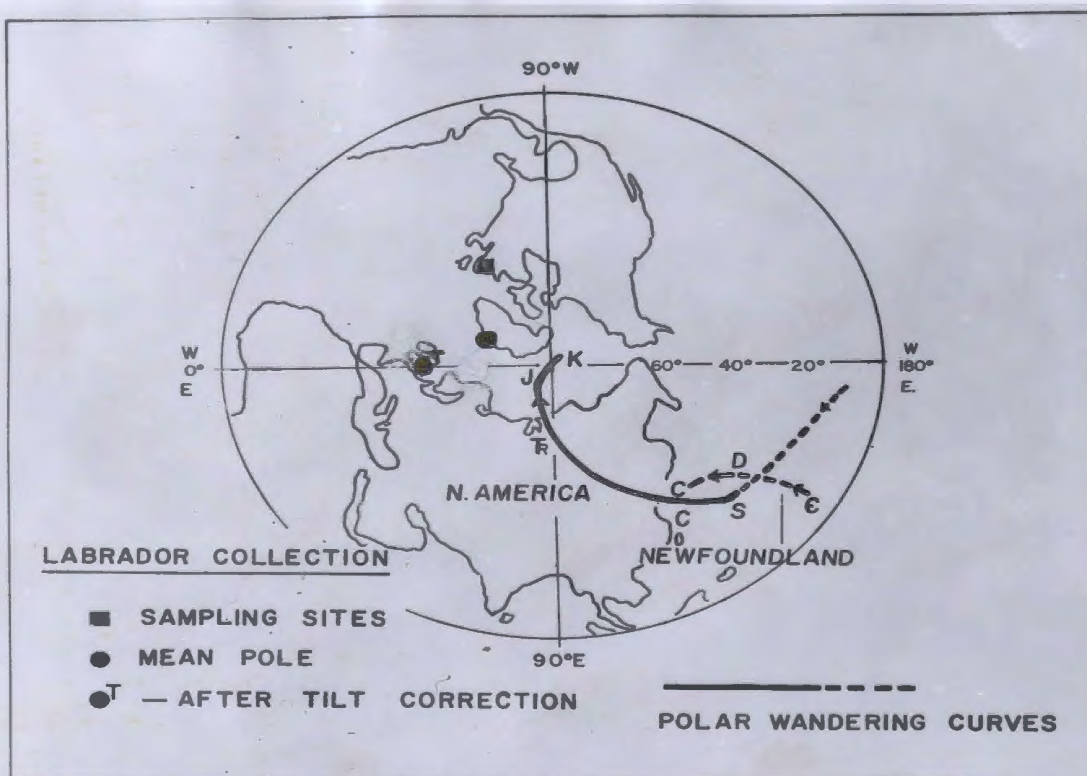
In addition to the individual means for the three sites, Figure 9 gives the mean magnetic direction of the basalt formation of Labrador. In Figure 9 the present geomagnetic field and the theoretical geocentric axial dipole field at Henley Harbour are also plotted. The significance of these directions will be discussed in Section 4.12.

4.11 Obtaining the palaeomagnetic pole

The usefulness of expressing the palaeomagnetic results in terms of the ancient magnetic pole was discussed in Section 3.5.3. The ancient position of the geomagnetic pole was obtained from the combined mean magnetic vector of the basalt formation at Henley Harbour and Table Head and are:

73.5°N latitude; 22.1°W longitude.

From the general similarity of the mean magnetization vector in the rocks with the present dipole direction in Labrador,



ANCIENT POLE POSITIONS FROM MAGNETISM OF NORTH AMERICAN ROCKS

Fig. 10

it is probable that the computed palaeomagnetic pole has the same polarity as the present magnetic pole in the northern hemisphere. The ancient pole position is plotted in Figure 10 along with the polar wandering curves relative to North America and Newfoundland based upon published results by various authors.

4.12 Stability of magnetization of basalt flows from Labrador

A partial indication of stability exists when the mean direction of magnetization of a suite of rocks is significantly different from that of the present field at the site. However, it can be seen from the above results (Fig. 9) that although the present field directions are outside the 95% circle of confidence, the observed mean direction is not very different from that of the present dipole field or of the actual field at the localities. This indicates that a significant unstable component may be contributing to the observed magnetization, and hence it becomes important to test for stability in one of two ways: either by showing that the measured directions of magnetization are after all consistent with stability, or by removing unstable components and remeasuring the magnetization of the rocks. Either result can be obtained from the same stability tests in the laboratory.

The relatively small angle between the existing field or the axial geocentric dipole field on the one hand, and the direction of the mean remanent vector on the other, is an indication that a viscous component due to the present earth's field may be prominent in the measured NRM. Hence this precludes the formulation of palaeomagnetic inferences from these results until complete stability tests have been carried out; these will consist in verifying the presence of any strong viscous component superposed on the primary component of magnetization and, if it is present, to remove it by means of either alternating field or thermal demagnetization. Instrumentation for both purposes is under construction at Memorial University; the alternating field demagnetization equipment was completed at the close of the author's investigations, and therefore only a partial stability analysis, applied to specimens from Henley Harbour and Table Head, could be carried out. The results, which are shown in Table 9 and Figure 11, and are discussed below, are therefore still incomplete.

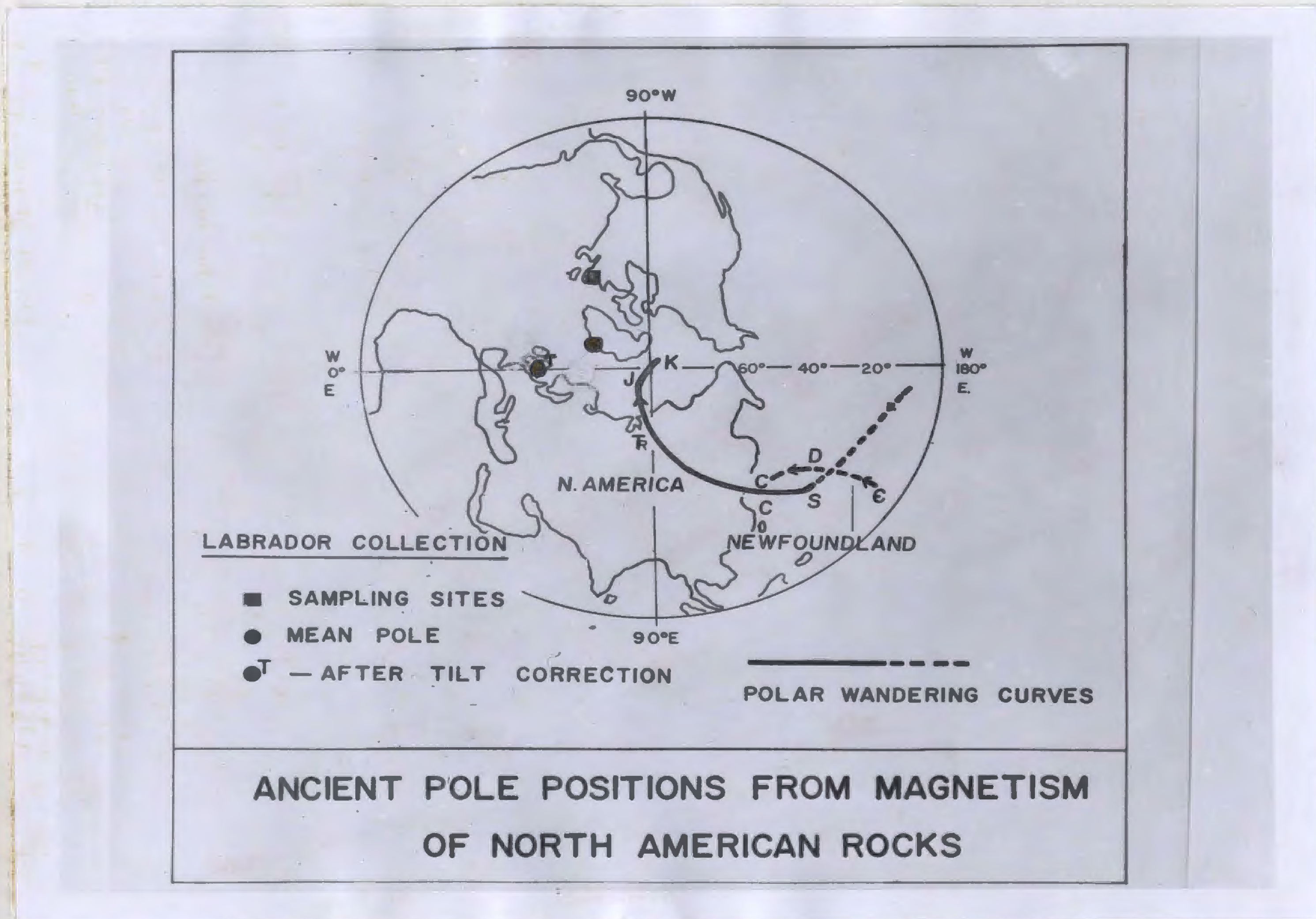
Four specimens from Henley Harbour and two from Table Head were selected for alternating field demagnetization test. They were successively treated at peak fields of 20, 50, 108, 243 and 324 oe, respectively, by G. W. Pearce of the Physics Department. The intensities, J, fell down gradually with increasing field (Fig. 11a), but except for one specimen (HH11AZ) the value of J after demagnetization at 324 oe was never less than 22% of its original value; this is an indication of stability (Doell and Cox, 1963). Moreover the change in direction of magnetization is not very large even after treatment in 345 oe field. If anything, the four Henley Harbour vectors are clustered approximately about the mean direction of all Henley Harbour basalts before demagnetization

($\bar{A}$ = 12.0°, $\bar{D}$ = +75.9°) and the two Table Head vectors are close to the mean direction of all Table Head basalts before demagnetization ($\bar{A}$ = 72.6°, $\bar{D}$ = +72.8°).

The demagnetization results, tabulated in Table 9a, indicate that the measured magnetization is largely due to a stable component although this may not be the primary component responsible for the original magnetization.

In addition to the above six specimens, which before demagnetization had exhibited NRM directions and intensities representative of the Henley Harbour and Table Head basalts as a whole, three specimens (HH1C3, HH21A4 and $HH8A_1$) which had an original anomalous direction of magnetization, were also given alternating field treatment; the results of these are shown in Table 9b and plotted in Figure 11b.

Thin-section studies were carried out by A. C. Nautiyal of the Geology Department on one sample each of the Henley Harbour and Table Head basalts and the studies indicate that the rocks are partly altered. In the Henley Harbour specimen, augite had mostly altered to chlorite and partly to deep red hematite, the latter mineral occurring in local concentrations. Zoisite (a variety of epidote) is very commonly distributed throughout the rock. On the whole, hematite is more prominent than magnetite, which is rather unusual for a basalt. The Table Head specimen studied shows less alteration, although in this case also, hematite, which is believed



ANCIENT POLE POSITIONS FROM MAGNETISM OF NORTH AMERICAN ROCKS

Fig. 10

TABLE 9A

REMANENT MAGNETIZATION OF SELECTED "REGULAR" BASALT SPECIMENS AFTER SUCCESSIVE ALTERNATING FIELD DEMAGNETIZATION

TREATMENT (Peak a.c. field, oe.)	MAGNETIZATION AFTER TREATMENT Azimuth (deg.)	Dip (deg.)	Intensity (e.m.u./cc $\times 10^{-4}$)
	HH 11 A2		
before demagnetization	18	68	6.1
108	20	53	1.4
243	28	57	0.72
345	308	54	0.34
	HH 16 B1		
before demagnetization	347	60	5.5
108	74	77	3.9
243	20	78	2.5
324	59	72	2.5
	HH 20 A3		
before demagnetization	27	75	14.7
108	10	71	8.8
243	4	77	8.6
324	17	70	6.0
	TH 2 B1		
before demagnetization	54	57	6.7
20	45	55	6.7
50	58	62	5.5
60	55	65	5.1
150	29	65	5.0
400	63	68	3.1
	TH 5 A4		
before demagnetization	49	71	13.0
108	59	71	6.4
243	64	73	4.3
324	77	69	2.9

TABLE 9A (cont'd)

REMANENT MAGNETIZATION OF SELECTED "REGULAR" BASALT SPECIMENS AFTER SUCCESSIVE ALTERNATING FIELD DEMAGNETIZATION

TREATMENT (Peak a.c. field, oe.)	MAGNETIZATION AFTER TREATMENT Azimuth (deg.)	Dip (deg.)	Intensity (e.m.u./cc $\times 10^{-4}$)
	HH 16 A1		
before demagnetization	78.3	76.6	10.1
20	57.1	77.3	9.2
50	65.7	76.6	7.4
100	69.6	76.9	6.6
282	84.9	74.8	4.8
324	352.2	80.4	

TABLE 9B

REMANENT MAGNETIZATION OF SELESTED "ANOMALOUS" BASALT SPECIMENS AFTER SUCCESSIVE ALTERNATING FIELD DEMAGNETIZATION

TREATMENT (Peak a.c. field, oe.)	MAGNETIZATION AFTER TREATMENT Azimuth (deg.)	Dip (deg.)	Intensity (e.m.u./cc $x10^{-4}$)
	HH 1 C3		
before demagnetization	131	21	3.4
20	127	43	2.9
50	150	45	1.2
100	130	75	0.95
	HH 8 A1		
before demagnetization	218	-65	3.3
20	195	-63	-
50	173	-72	2.5
100	141	-74	2.6
150	151	-75	2.7
	HH 21 A4		
before demagnetization	198	73	1.3
27	193	72	0.84
50	181	73	0.33
100	187	42	0.05
150	352	-11	0.28
300	132	32	0.14

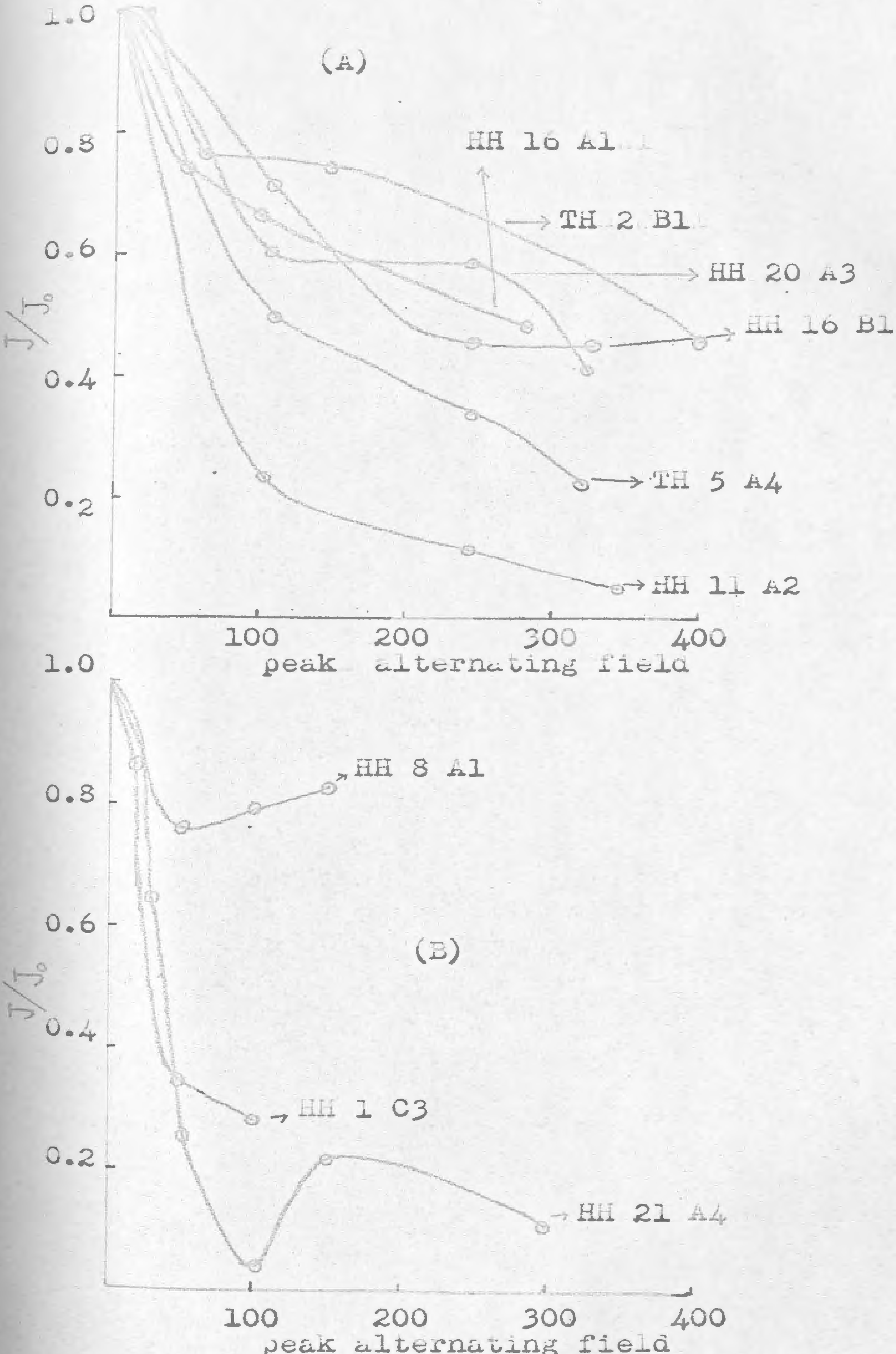


Fig. 11 VARIATION OF J/J_o WITH THE DEMAGNETIZING FIELD FOR CERTAIN SELECTED BASALT SPECIMENS FROM LABRADOR
(A) REGULAR (B) ANOMALOUS

demagnetization could remove a magnetization of low-temperature origin due to hematite without affecting the postulated high temperature TRM residing in magnetite. Until these additional stability tests, along with further thin-section studies have been carried out, the mechanism suggested above must remain tentative.

4.13 Significance of the pole position

The pole position obtained from the present data (73.5°N lat., 22.1°W long.) lies very far from the entire polar wandering curve relative to North America (Fig. 10). Hence it cannot be correlated either with a lower Palaeozoic pole, or with "typical" North American pole position relative to any subsequent geological period. This tends to make the rocks more suspect of possessing a major unstable component. The inferred position of the pole in the eastern Atlantic Ocean may indicate partial instability, in the sense that the magnetization vectors could have been pulled around part of the way from a position along the North American polar wandering curve to that of the present pole. A second alternative is that the high present latitude of the ancient pole may indicate that the rock formations are indeed younger than is believed to be the case by most geologists, but this suggestion must be taken with strong reservation, pending the confirmation of the remanent magnetization directions by more complete stability tests. Even if the latter test were to prove positive one would still have to explain why the palaeomagnetic pole, though in high latitude, lies rather far from the polar wandering curve relative to

North America.

4.14 Application of tilt corrections to the Henley Harbour data

The strike and dip of the Henley Harbour basalts has been estimated to be 20° and 14° SSE respectively. On the assumption that tilting of the formation was subsequent to its acquisition of remanent magnetism, the measured magnetic vector has been transformed relative to the original horizontal at the time of eruption. The extent to which this assumption is justified will depend on the results of more exhaustive stability tests. However, on structural grounds, the application of tilt corrections is unobjectionable here as there is no evidence that the beds have undergone rotation about a vertical axis. The transformed vector has the following direction (for Henley Harbour only):

$$\text{Azimuth, } \overline{A} = 60.1^{\circ}\,, \qquad \text{Dip, } \overline{D} = 71.7^{\circ}$$

It is of interest to note that the tilt correction has brought the transformed vector corresponding to the Henley Harbour data very close to the mean magnetic vector for the Table Head data (Fig. 9) where a tilt correction has not been made since the beds are within 3° or so of being horizontal. This indicates that the basalt formations from both sites were magnetized in fields having the same direction. Further it also indicates <u>partial</u> magnetic stability, as one would otherwise have expected closer alignment between the Henley Harbour and Table Head vectors <u>before</u> tilt correction. The ancient pole position has also been calculated from the corrected data and has the geographical co-ordinates

$$55.7^{\circ}\text{N latitude}, \qquad 2.2^{\circ}\text{E longitude}$$

This position of the pole is also plotted in Figure 10. The latter pole position is somewhat more significant than the uncorrected position, if it can be assumed that the magnetization was acquired prior to tilting of rocks.

CHAPTER 5

PRELIMINARY PALAEOMAGNETIC MEASUREMENTS OF THE MAGNETITE-RICH ARGILLITES FROM THE AREA NEAR SPRINGDALE, NEWFOUNDLAND

5.1 Preliminary remarks

During the summer of 1965, samples of "ferruginous argillite" were collected by the author from four sites at the west coast of Notre Dame Bay, to the northeast of Springdale, Newfoundland. Palaeomagnetic directions for some of these rocks were obtained with the astatic magnetometer, and the data, along with results from alternating field demagnetization tests, are presented and discussed in this chapter.

5.2 General considerations and geology

The present study constitutes the first stage of a palaeomagnetic contribution towards the "Whalesback Project" supported by the Geological Survey of Canada, in which the aim is to gain a better understanding of the geology in the general area of the British Newfoundland Exploration Company's Whalesback mine near Springdale.

In this project the rocks of primary interest are of Ordovician age. While most of the rocks in this area have been dated to the extent of distinguishing between a Silurian and an Ordovician origin, (and occasionally more closely, e.g. lower, middle or upper Ordovician) a major problem concerns their geological structure, as the rocks are often structurally deformed in such a way that geological dips and strikes can change drastically even over short distances; hence the orientation of the beds prior to folding frequently cannot be ascertained by standard geological procedures. Palaeomagnetism can help to solve such problems, provided the following two steps are taken:

(1) Palaeomagnetic pole positions for the time span concerned (in this case preferably a known sub-division of the Ordovician) must be first established relative to the region of interest, using palaeomagnetically suitable rocks. The rocks also must be relatively unmetamorphosed, magnetically stable, and positioned in such a way that a "tilt" correction can be applied to relocate them in the attitudes they assumed when they were magnetized.

(2) The ancient magnetic field inferred from (1) is then taken to be also the ambient field at the time of formation of other rocks in the same region whose contemporaneity with those under (1) has been established, but whose original position is unknown relative to azimuth (and perhaps also with regard to distinguishing the original "top" and "bottom" surfaces of the beds). If the remanent magnetization of these rocks can be reliably measured relative to their present attitude, they may then be "relocated" in their ancient position through alignment of the remanent vector (regarded as fixed within the rock) with the previously established ancient field direction.

The present studies were initiated as a part of Step (1) mentioned above. Although it is fairly difficult to find a rock formation to serve as a control, Dr. V. S. Papezik of the Geology Department at Memorial University suggested that the ferruginous argillites exposed in the area north of Springdale (Papezik, 1963) might prove suitable for that purpose; accordingly the initial collections have been made at the argillite exposures. These rocks grade from strongly magnetic, magnetite-bearing, grey argillites to red argillites whose colour is probably derived largely from hematite.

Unfortunately, the out-crops are often very small, making it difficult to obtain a statistically significant number of samples at a given site. On the other hand, petrologically similar rocks of the same age are exposed in a fairly large number of places not only in the Springdale area, but also in other parts of north-central Newfoundland. Under the circumstances, it is best to sample as large a number of contemporaneous out-crops in this general region, in order to establish Ordovician pole positions with respect to it. For this reason, reddish argillites as well as some volcanic rocks of Ordovician age were also sampled from the area north of Botwood, 70-75km south-east of Springdale. A difficulty is that the distinction between the "top" and "bottom" faces of the beds cannot always be made, and the assumption that the beds have not rotated significantly about a vertical axis is not conclusive in the collections at any particular site.

However, if magnetic stability can be demonstrated, the results should yield at least a reasonably constant value for the angle between the remanent vector and the bedding plane; i.e. the least that can be expected from such a study is that it will furnish a consistent Ordovician magnetic dip. Those sites where the "top" vs "bottom" orientation is known should, in addition, determine the sense of the magnetization vector, i.e. determine whether the dips are upward or downward. The most that can be expected is that complete Ordovician palaeomagnetic directions (magnetic azimuths as well as the dips) may be inferred on a statistical basis. i.e., if the azimuths of the rocks at the various sites after correction for tilt show a preferred alignment, then the mean direction of magnetization (with all sites given equal weight) might be taken as the most probable field direction at the time the rocks were formed, with a confidence that would depend on the number of sites and the scatter of directions between them. Conversely, the tilt correction is more likely to be justified when it results in a significant alignment of the mean remanent vectors between sites. Contrary to the case of the dips, the confirmation of relative consistency between azimuths at different sites has to await completion of measurements from a large number of sites.

For the present initial study, only some argillite samples from four sites (all in the Springdale area) have been examined. Samples from various exposures of argillite and volcanic rock collected in 1965 north of Botwood, and also from a further collection of strongly magnetic argillites in the Springdale area by members of the Physics Department in 1966, have been left for further study. Hence conclusions presented in this chapter can be only tentative.

The localities of the four collection sites are:

Locality	Latitude	Longitude	No. of Samples
Nicky's Nose	49°45'N	55°56'W	3
Little Bay	49°40'N	55°56'W	2
Silverdale	49°45'N	55°56'W	3
Tilt Cove	49°50'N	55°45'W	2

Nicky's Nose:

Magnetite-rich argillites are exposed at Nicky's Nose as very small out-crops on a steep hill. The argillite is thin-bedded, dark and probably minutely drag-folded; the rocks are also strongly magnetic and relatively unaltered. Papezik (1963) reports that the underlying rock is basalt. Accoring to MacLean (1947), the argillites in this area are of early Ordovician age. The dips are larger than 80° and the tops and bottoms of the beds are not distinguishable.

Little Bay:

Near the village of Little Bay, samples were taken from the edge of a pond. The out-crop is again very small, and the rocks are similar to those at Nicky's Nose but reddish rather than grey in colour and less strongly magnetic. Some of the beds show distinct but uneven banding and, in certain places, the bands are green, which may be due to epidote, indicating some degree of metamorphism. The rocks were found to be dipping at very steep angles ($>85^{\circ}$) but the tops and bottoms of the beds are not distinguishable. The long axis of the out-crop is approximately along strike.

Silverdale:

The out-crop of "ferruginous argillite" at Silverdale is similar petrologically to that at Nicky's Nose, but again the samples proved to be less strongly magnetic. The out-crop is located on the coast at Little Bay and is fully above sea-level only at low-tide. The argillites were found to contain some epidote and occasional bands of jasper. The dips again are very steep. Tops and bottoms were undistinguishable at one of two sub-sites, but could be distinguished at the other sub-site because of graded bedding.

Tilt Cove:

At Tilt Cove, about 50km north-east of Springdale, the argillite was sampled inside the Tilt Cove copper mine and in the vicinity of the mine. The formations appeared to be hematitic from appearance (red-coloured and fine-grained). Sample TC1 was collected inside the mine at the 2000-ft. level below the surface.

The dips are steep, but the upper and lower faces of the beds can be distinguished by means of sorting of grains. Several other samples were obtained from a small, patchy out-crop on a hill above the mine. The geological dip here is less steep, and the tops and bottoms of the beds can again be distinguished by grain sorting. So far measurements have been made only on three specimens of one sample from Tilt Cove (TC1, collected inside the mine).

Because the rocks generally showed excellent banding, accurate strike and dip values were easy to obtain. The measurements were made close to the samples and are listed below (where the word "tilt" will be used here instead of "dip", to avoid confusion with the magnetic dip).

Sample No.	Strike	Geological Tilt (or Dip)
NN1	300°	98°NE (or 82°SW)
NN2	301°	90°NE (or 90°SW)
NN3	301°	92°NE (or 88°SW)
LB2	18°	90°E (or 90° W)
LB3	18°	87½°E (or 92½°W)
SD1	291°	100°N (or 80°S)
SD2	291°	100°N (or 80°S)
SD4	297½°	83°NE
TC1	256°	89°N
TC3*	242½°	68½°NW

*Not measured

It may be noted that alternative values of tilt and tilt direction have been given in the case of those samples where tops and bottoms of the beds are not distinguishable.

5.3 Results

The remanent magnetization of a total number of 16 specimens was measured with an astatic magnetometer. Individual specimen directions and intensities of magnetization are tabulated in Table 10, and sample mean values in Table 11. These results, which are with respect to the present horizontal plane, do not have any palaeomagnetic significance, except in the sense that the attitudes of the remanence vectors relative to the present horizontal plane must be considered in any assessment of stability; in general these have fairly steep dips and could have been deflected towards the earth's field direction in relatively recent times. While the number of samples so far examined is too small to permit any statistical significance test of the deviation between the remanent magnetization direction in these rocks and that of the present field (or the axial geocentric dipole field), some of the specimens have been subjected to alternating field demagnetization, and the results will be discussed in Section 5.4.

Assuming for the present that magnetic stability can be demonstrated, the NRM directions have palaeomagnetic significance only if they are corrected for the tilt of the bedding plane. The question whether or not such corrections are applicable at the Springdale sites has been discussed in Section 5.2; in any case,

TABLE 10

REMANENT MAGNETIZATION OF THE INDIVIDUAL SPECIMENS FROM ROCKS IN THE SPRINGDALE AREA

SPECIMEN NO.	AZIMUTH(A) (degrees)	DIP(D) (degrees)	INTENSITY OF MAGNETIZATION(J) (e.m.u./cc x 10^{-4})
NN1 A1	3	+70	366
NN2 A1	15	41	23.6
NN2 B1	59	54	19.0
NN3 A1	63	57	9.7
NN3 B1	63	62	18.3
LB2 A3	268	+48	0.94
LB2 B1	290	77	0.71
LB2 B2	312	69	0.27
LB2 B3	102	35	0.91
LB3 A1	228	63	0.53
LB3 B1	164	53	0.26
SD1 A1#	(64)	(+84)	-
SD2 A1#	(23)	(87)	-
SD4 A1#	(273)	(80)	-
TC1 A1	37	+70	0.68
TC1 A2	60	65	0.84
TC1 A3	86	74	0.84

Inhomogeneous magnetization; quoted values of A and D are doubtful.

the adjustment is a major one in view of the fact that all geological tilts and most of the magnetic dips are quite steep; therefore, tilting of the beds into the horizontal plane corresponds to a rotation of the magnetization vectors through large angles into directions with low dip. In the cases where tops and bottoms of the beds are not distinguishable, they can be rotated about strike in either sense, resulting in two alternative NRM directions after tilt correction.

The corrected results for samples and the mean results for each site are tabulated in Tables 12 and 13, respectively, with both alternative directions being shown where applicable. The results in Table 12 have also been plotted on an equal-area net in Fig. 13.

Palaeomagnetic pole positions corresponding to the data of Table 13 are shown in Table 14, but again in view of the small number of samples, these have virtually no statistical significance and are presented only for comparison with better-established pole positions for the lower Palaeozoic. Relative to North America, however, reports of the latter in the literature are also scarce, and Irving (1964) lists only one reliable Ordovician pole: this was inferred from the upper Ordovician Juniata formation in Pennsylvania by Collinson and Runcorn (1960) and has the co-ordinates: 20°N, 153°E. The pole positions in Table 14 are quite scattered, though the fact that they are also in low present latitudes, is probably significant. Nevertheless the longitudes inferred from the present samples differ substantially from that quoted by

TABLE 11

MEAN SAMPLE MAGNETIZATION WITH RESPECT TO TRUE NORTH AND PRESENT HORIZONTAL *

SAMPLE NUMBER	NUMBER OF SPECIMENS AVERAGED	MEAN AZIMUTH $\bar{A}$ (deg)	MEAN DIP $\bar{D}$ (deg)	MEAN INTENSITY OF MAGNETIZATION (emu/cc x 10^{-4})
NN 1	1	3.0	70.0	366
NN 2	2	8.8	47.6	21.3
NN 3	2	62.9	59.6	14.0
LB 2	4	292.2	83.6	0.71
LB 3	2	190.8	61.9	0.40
TC 1	3	58.9	70.6	0.78

* Silverdale samples omitted because of inhomogeneity

TABLE 12

MEAN SAMPLE DIRECTIONS WITH RESPECT TO HORIZONTAL BEDDING PLANE AND TRUE NORTH, AFTER ALTERNATIVE TILT CORRECTIONS

Sample No.	Azimuth (degrees)	Dip (degrees)	Alternative	Geological Tilt Assumed
NN1	19.1	-27.6	1	98° NE
	220.9	27.6	2	82° SW
NN2	11.9	-38.8	1	90° NE
	230.1	38.8	2	90° SW
NN3	48.5	-27.6	1	92° NE
	193.5	27.6	2	88° SW
LB 2	292.6	-6.3	1	90° W
	103.4	6.3	2	90° E
LB 3	260.0	1.2	1	92.5° W
	135.9	-1.2	2	87.5° E
TC 1	4.5	-4.8	*	89° N

* Tilt is unambiguously given

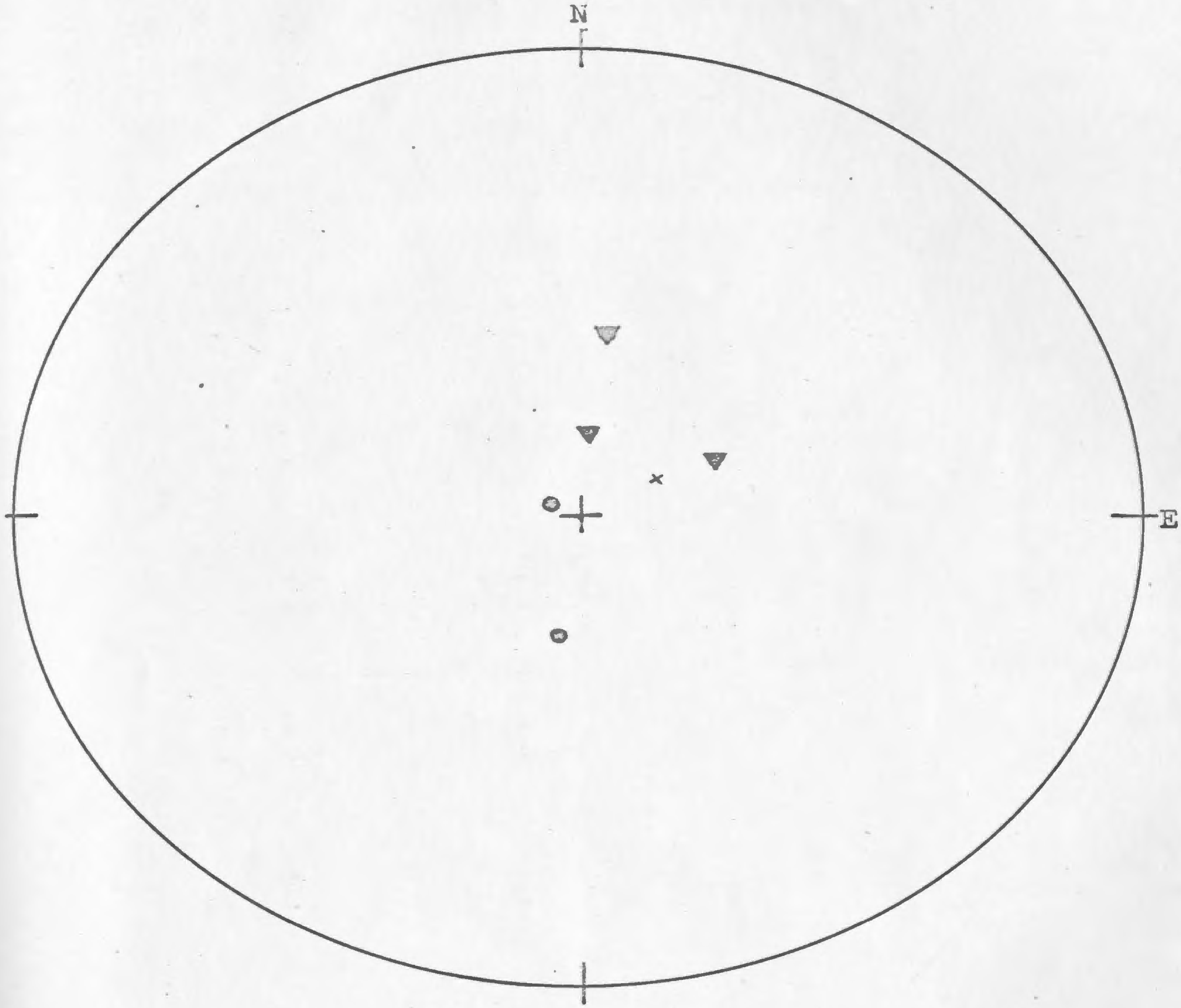


Fig. 12a. MEAN DIRECTIONS OF MAGNETIZATION OF THE ARGILLITE SAMPLES FROM SPRINGDALE AREA

Equal area projection, the primitive being the present horizontal

North poles all downward

▽ Samples from Nicky's Nose
⊙ Samples from Little Bay
✕ Samples from Tilt Cove

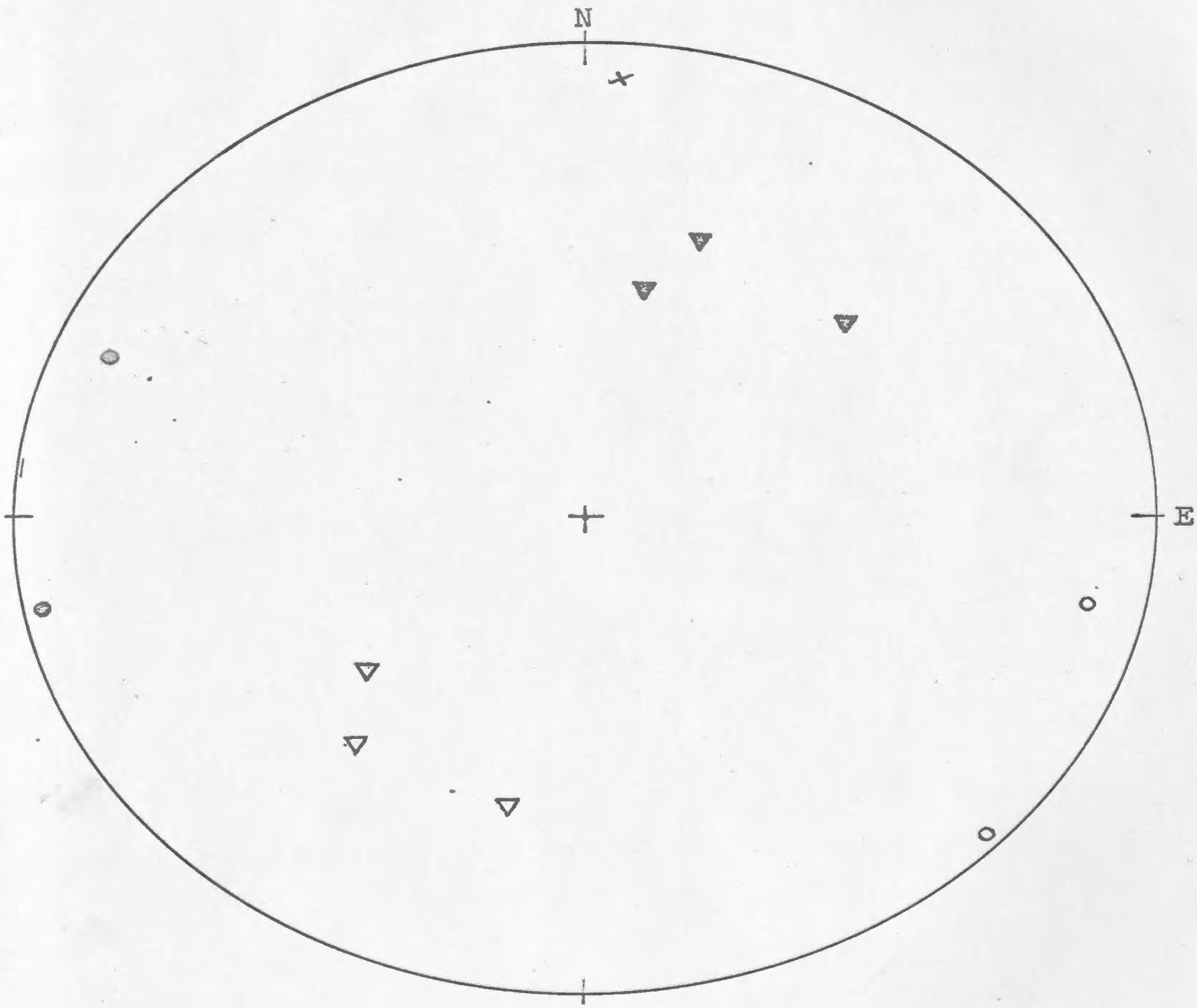


Fig. 12b. MEAN DIRECTIONS OF MAGNETIZATION OF THE ARGILLITE SAMPLES FROM SPRINGDALE AREA

Equal area projection, the primitive being the horizontal bedding plane

Solid symbols indicate north poles upward
Hollow symbols indicate north poles downward

▽ ▽ Nicky's Nose, Alternatives 1 and 2
● o Little Bay, Alternatives 1 and 2
× Tilt Cove

(see text)

TABLE 13

MEAN SITE DIRECTIONS OF MAGNETIZATION WITH RESPECT TO HORIZONTAL BEDDING PLANE AND TRUE NORTH

SITE	NUMBER OF SAMPLES AVERAGED	MEAN AZIMUTH (degrees)	MEAN DIP (degrees)	ALTERNATIVE
Nicky's Nose	3	27.0	-32.2	1
		220.3	31.0	2
Little Bay	2	276.3	-2.7	1
		119.7	2.7	2
Tilt Cove	1	4.5	-4.8	unambiguous

TABLE 14

PALAEOMAGNETIC POLE POSITIONS OBTAINED FROM THE DATA OF THE ROCKS FROM THE SPRINGDALE AREA (approximate values only)

SITE	POLE POSITION LATITUDE	POLE POSITION LONGITUDE	ALTERNATIVE
Nicky's Nose	19°N	97°E	1
	11°N	76°E	2
Little Bay	3°N	40°E	1
	18°N	170°W	2
Tilt Cove	38°N	119°E	unambiguous

* All pole positions have been transposed to the northern hemisphere.

Collinson and Runcorn.

5.4 <u>Assessment of stability</u>

The palaeomagnetic study of these argillites is still incomplete and the results do not yet lend themselves to a final interpretation. This is not principally because of the ambiguity (in most cases) with regard to top vs bottom faces of the beds, because, as shown by the data in Tables 12, 13 and 14, this affects not so much the <u>position</u> of the palaeomagnetic axis, as its <u>polarity</u>. In all cases this axis after tilt correction, makes a large angle with the present spin and magnetic axes of the earth, resulting in poles at low latitude which are qualitatively consistent with the generally low latitudes of lower Palaeozoic pole positions relative to North America reported in the literature. The uncertainty in the present case is due rather to the small number of samples so far examined, and to the fact that the magnetization vectors relative to the <u>present</u> horizontal plane make relatively small angles with the present spin or magnetic axes, so that independent stability tests assume special importance.

The fact that the magnetization vectors in individual specimens from the same sample generally show a misalignment considerably in excess of experimental error, is an indication of at least partial instability, as it cannot be explained by deformation of the rock. Similarly, there are relatively large differences between mean sample directions at the same site, as well as between sites both before and after tilt correction (Tables 11 and 12).

In order to randomize the magnetization directions due to the possible presence of a soft component, and thereby cause greater alignment between these directions, certain selected specimens have been successively demagnetized in peak fields of 108, 202, 324 and 405 oe respectively. The tilt correction is incorporated in the results of Table 15 giving the direction of magnetization observed after each successive alternating field treatment. The demagnetization results are plotted in Figures 13a, ~~b~~, and b. Although after the alternating field treatment, the scatter is sometimes less and in no case substantially more than before treatment, the fact that few specimens have been tested do not permit the estimate of an optimum field of demagnetization. However, changes in the directions of magnetization produced by the stability treatment, have been sufficiently small in all cases to make it safe to deduce the presence of a significant hard component of magnetization.

Considerable scatter in the positions of the pole from different sites may also be due to the uncertainty of the strikes and dips of the beds. Moreover, great use has been made of the "tilt corrections", the results of which in this case are decisive in interpretation, though as stated in Section 5.2, the geological control needed for applying these tilt corrections is lacking at present.

However, the result that the pole positions are in low latitudes, together with the demonstration of at least partial stability against demagnetizing fields up to 400oe, are promising

TABLE 15

REMANENT MAGNETIZATION OF SELECTED ARGILLITE SPECIMENS AFTER SUCCESSIVE ALTERNATING FIELD DEMAGNETIZATION

TREATMENT (Peak a.c. field, oe.)	MAGNETIZATION AFTER TREATMENT Azimuth (deg.)	Dip (deg.)	Intensity (e.m.u./cc)
	NN1A1		
before demagnetization	3	70	36.6×10^{-3}
108	200	88	3.8 "
202	263	83	2.8 "
324	297	75	1.1 "
405	110	69	1.6 "
	NN2A1		
before demagnetization	18	41	2.36×10^{-3}
108	341	71	0.97 "
202	353	70	0.61 "
324	317	75	0.50 "
405	357	47	0.54 "
	NN3A1		
before demagnetization	63	62	1.83×10^{-3}
202	23	68	0.35 "
324	14	49	0.27 "
405	0	31	0.21 "
	TC1A2		
before demagnetization	60	65	8.4×10^{-5}
108	196	72	4.5 "
	LB2A3		
before demagnetization	269	48	9.4×10^{-5}
108	45	82	4.1 "

* Azimuth and dip are relative to the present horizontal and true north

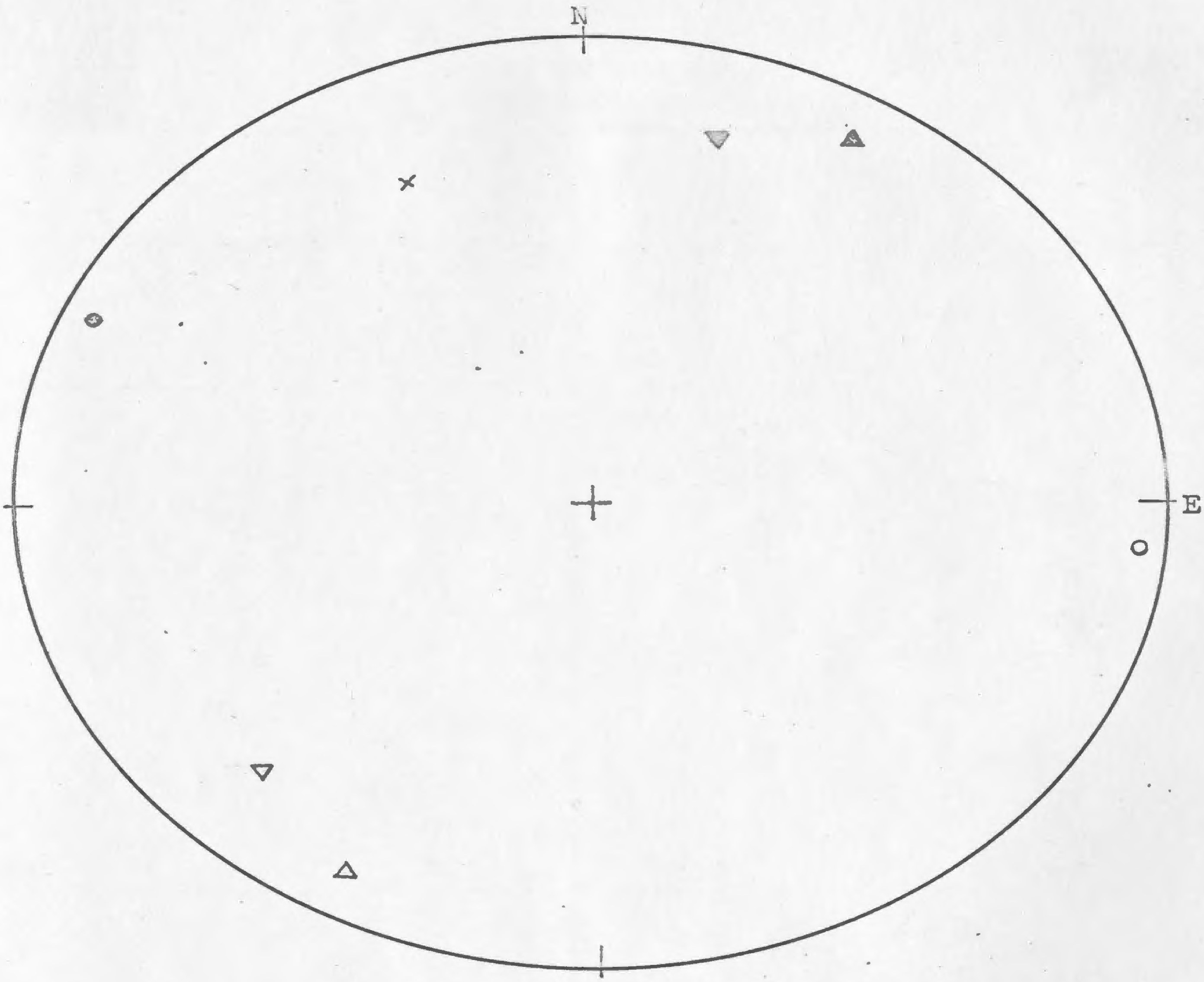


Fig. 13a DIRECTIONS OF MAGNETIZATION OF CERTAIN SELECTED ARGILLITE SPECIMENS AFTER DEMAGNETIZATION IN 108 OE FIELD

Equal area projection, the primitive being the horizontal bedding plane

▲ △ Specimen NN 1 A1, Alternatives 1 & 2
▼ ▽ Specimen NN 2 A1, Alternatives 1 & 2
● ○ Specimen LB 2 A3, Alternatives 1 & 2
× Specimen TC 1 A3

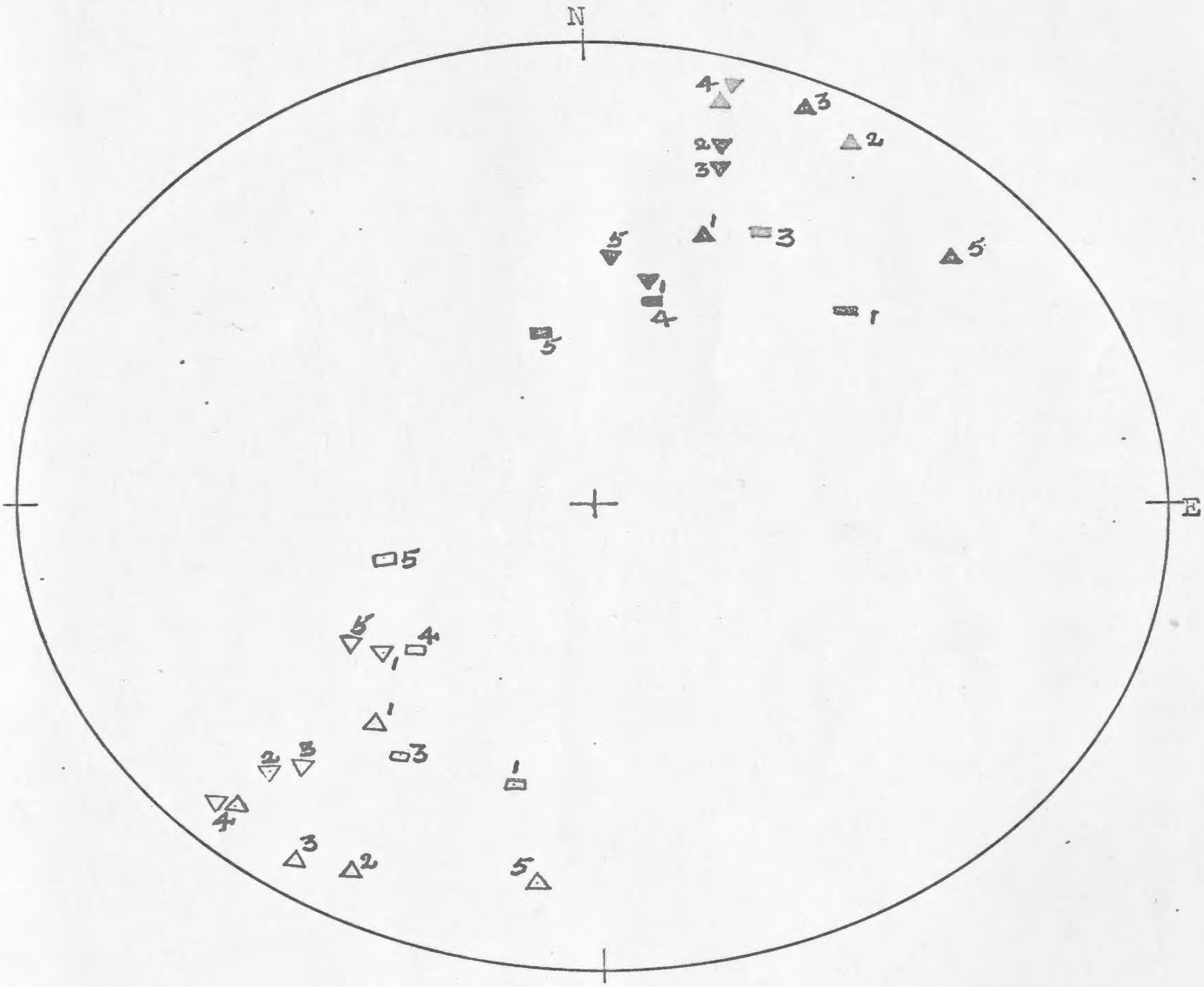


Fig. 13b. Equal area plot showing variation in directions of magnetization of certain selected argillite specimens from Nicky's Nose. Solid symbols indicate north poles upward and hollow symbols indicate north poles downward. The primitive is the horizontal bedding plane. Numbers 1-5 indicate the corresponding magnetization directions before demagnetization, and after 108, 202, 324, and 405 oe. alternating field demagnetization respectively.

▲ △ NN 1 Al; Alternatives 1 & 2
▼ ▽ NN 2 Al, Alternatives 1 & 2
▬ ▭ NN 3 Al, Alternatives 1 & 2

criteria, indicating that future systematic studies of more samples of argillites from additional sites in north-central Newfoundland should establish a reliable Ordovician pole relative to this region.

CHAPTER 6

SOME PALAEOMAGNETIC MEASUREMENTS ON SAMPLES FROM THE GRAND BANKS OF NEWFOUNDLAND

6.1 Preliminary remarks

Results of a preliminary palaeomagnetic study of the Virgin Rocks shoals on the Grand Banks of Newfoundland are reported in this chapter. The Virgin Rocks specimens were cut in the form of cubes, and as such a different procedure of measurement has been adopted. The results are discussed with reference to the earlier palaeomagnetic work in Newfoundland.

6.2 General considerations

Reports of palaeomagnetic studies of underwater samples are comparatively scarce, though recently Harrison (1966) published the results of magnetic measurements on oriented deep-sea sediments from the Pacific Ocean obtained by standard coring techniques. However, only unconsolidated material is recovered in this way, whereas no results based upon oriented bedrock samples (either in the form of irregular pieces, or of drill cores) appear to have been published.

On the other hand the continental shelf is accessible to aqualung divers to a depth of 50m. or more, which opens up the prospect that bedrock samples from exposures on the shelf could be oriented insitu and recovered by divers using techniques similar to those applied to the collection of oriented samples on land. This was shown to be feasible during a reconnaissance geological expedition to the Virgin Rocks shoals on the Grand Banks, about 170km. east of the Avalon Peninsula, conducted by the Memorial University in 1965 (Deutsch and Lilly, to be published). The rocks were oriented under water with the aid of a pressurized Brunton compass, dislodged by explosive and brought to the surface by the divers. However, only three samples suitable for palaeomagnetic measurement were recovered: these were from two shoals separated by about 2km. and having respective minimum depths of 9 fathoms (16.5m.) and 17 fathoms (31m.). The locality is at Latitude 46°25'N, Longitude 50°49'W (Fig. 14). Some details are given below:

Sample No.	No. of Specimens Cut	No. of Specimens Measured	Minimum Depth to Shoal (fathoms)	Geological Structure at Shoal Strike	Geological Structure at Shoal Tilt
VR 1	2	2	9	113°	10°S
VR 2	5	5	17	5°	29°E
VR 3	8*	3	17	5°	29°E

* Five cubes were inhomogeneously magnetized.

6.3 Geology

The first direct underwater geological exploration of the Virgin Rocks area was reported by Lilly (1965). He described the material of the shoals as tough, well-indurated, greenish-grey, fine-

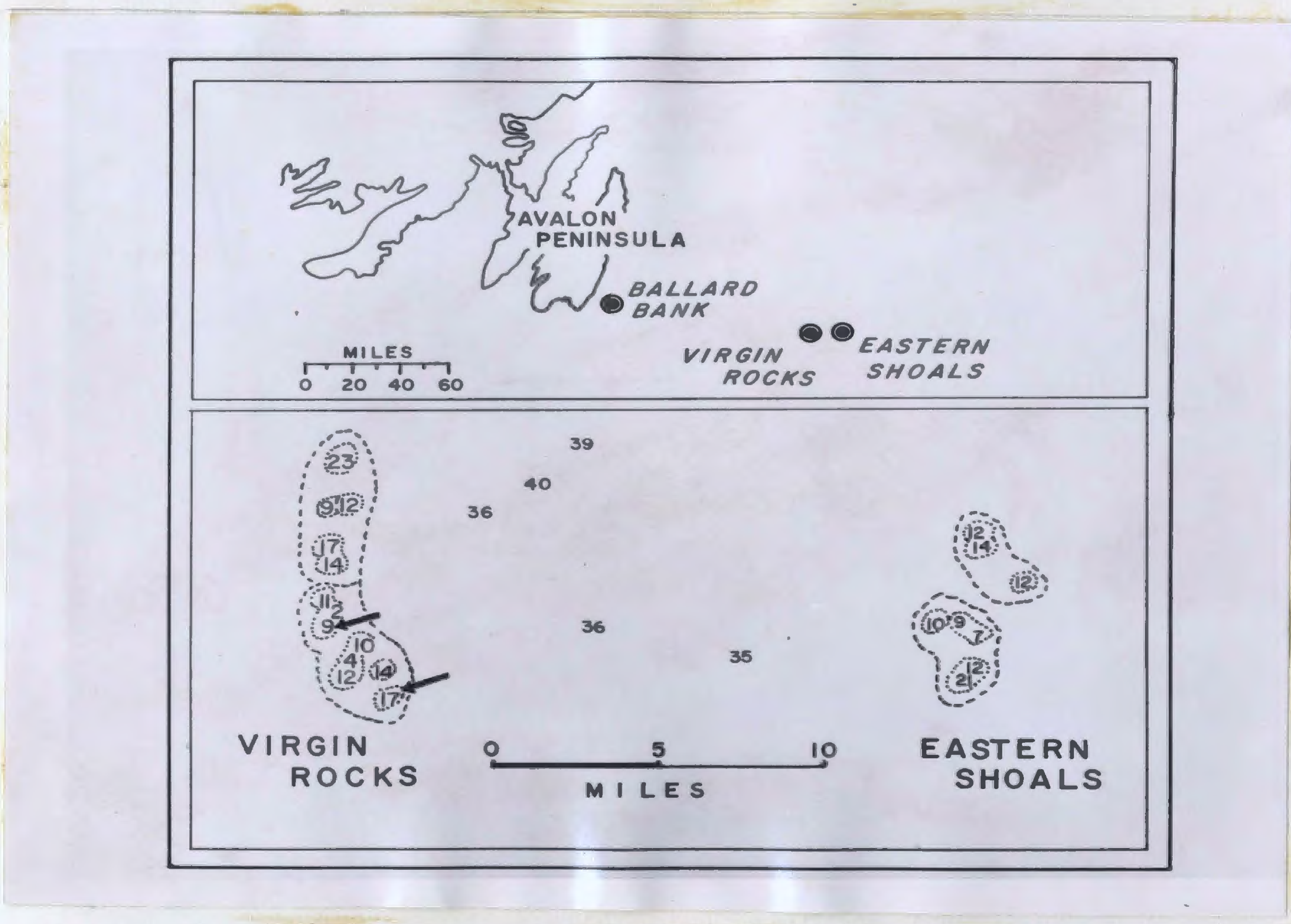


Fig. 14 LOCALITY OF THE VIRGIN ROCK SHOALS
(reproduced from Deutsch and Lilly, to be published)

grained microbreccia, containing fragments of quartz, granite gneiss, granite, diorite, feldspar, mafic volcanic rock and zircon in the range of 1-3mm. in diameter, in a fine siliceous matrix. The coarsest breccias were found to contain scattered cobbles of granite up to 100mm. in diameter. Small grains of magnetite, which are fairly common though irregularly distributed, are probably responsible for most of the remanent magnetism exhibited by these rocks. The shoals are believed to be of late Precambrian age because of their lithological similarity to rocks of the late Proterozoic group outcropping on the Avalon Peninsula of eastern Newfoundland.

6.4 Procedures for measurement of the magnetic vector

The measurement of magnetization directions for a cubical specimen is slightly more complicated than that of a cylindrical specimen. The procedure is outlined below:

As in the case of the previous investigations, the samples brought from the field are set in plaster of Paris in the laboratory prior to the cutting of specimens. Cubes having 2.0-cm. edges were cut with a diamond saw and, although contamination with minute ferromagnetic chips from the steel blade was probably negligible, the surfaces of the cubical specimens were rubbed with emery paper as a precaution prior to measurement. Each cube was cut with one pair of its side faces parallel to the north-south direction and its top and bottom faces representing the insitu horizontal plane of the sample. The following notation was used:

+x, +y and +z are co-ordinates in the north, east and downward directions, respectively;

X'_1 is the cube face perpendicular to and pierced by the +x-axis;

Y'_1 is perpendicular to and pierced by the +y-axis;

Z'_1 is perpendicular to and pierced by the +z-axis (i.e. the bottom face of the cube);

and X'_2, Y'_2 and Z'_2 are faces opposite to X'_1, Y'_1 and Z'_1, respectively.

Also, let X_1, Y_1, Z_1, X_2, Y_2 and Z_2 represent the scale deflections (usually in mms.) observed when X'_1, Y'_1, Z'_1, X'_2, Y'_2 and Z'_2, respectively, are facing east.

The magnetometer has been set up with the north-seeking end of the lower magnet pointing north and its south pole facing the observer. All measurements are referred to a standard orientation, which is that of the relevant cube face facing east; then any magnetization vector parallel to the east-west direction (i.e. perpendicular to the east and west faces) and having its north pole in the east direction will cause a maximum deflection of the astatic system, the movement of the light spot being to the right. The measurement procedure was as follows:

Step I. Insert the cube in the specimen holder with the Z'_2 and X'_1 faces facing top and east respectively, so that the Y'_1 face is towards the observer (position I in Fig. 15). Take the reading X_1.

Rotate the cube clockwise by 90^o about the vertical axis, Y'_2 facing east. Take the reading Y_2. Repeat this procedure with two further clockwise rotations by 90^o, obtaining the readings X_2 and Y_1.

Rotate the cube clockwise by another 90^o, bringing it back to its original position, with X'_1 facing east. If the reading differs from the first value of X_1, the difference is attributed to drift and is distributed over the four readings, X_1, Y_2, X_2 and Y_1. (No further reference will be made to this drift correction, which is applied also in the case of Steps II - VI, below, based always on a repeat measurement at the end of the step).

Step II. Rotate the cube by 180^o about the horizontal axis so that its position corresponds to position 2 of Fig. 15. Take the reading X_1.

Rotate the cube anticlockwise by 90^o about the vertical axis and measure Y_2. Repeat this procedure with two further anticlockwise rotations by 90^o, obtaining X_2 and Y_1.

Step III. Re-insert the cube as shown in position 3, Fig. 15 and measure Y_1. Rotate clockwise about the vertical axis in 90^o steps as before and read Z_2, Y_2, Z_1, and Y_1, respectively.

Step IV. Rotate the cube by 180^o about the horizontal axis so that position 4 of Fig. 15 is obtained, and measure Y_1. Rotate anticlockwise about the vertical axis in 90^o steps and measure Z_2, Y_2,

Z_1, and Y_1, respectively.

Step V. Re-insert the cube in the specimen holder as shown in position 5, Fig. 15, and measure Z_1. Rotate clockwise about the vertical axis in 90^o steps and obtain X_2, Z_2, X_1, and Z_1, respectively.

Step VI. Rotate about the horizontal axis by 180^o so that position 6, Fig. 15, is obtained, and measure Z_1. Rotate anticlockwise about the vertical axis in 90^o steps and measure X_2, Z_2, X_1, and Z_1, respectively.

In this way four readings are obtained for the east orientation of each cube face (not counting repeat readings for drift correction), yielding a total of 24 measurements. The procedure is summarized below:

Starting Position in Fig. 15	1	2	3	4	5	6
Sense of rotation C - clockwise A - anticlockwise	C	A	C	A	C	A
1st reading in Set	X_1	X_1	Y_1	Y_1	Z_1	Z_1
2nd reading in Set	Y_2	Y_2	Z_2	Z_2	X_2	X_2
3rd reading in Set	X_2	X_2	Y_2	Y_2	Z_2	Z_2
4th reading in Set	Y_1	Y_1	Z_1	Z_1	X_1	X_1
1st (repeat)	X_1	X_1	Y_1	Y_1	Z_1	Z_1

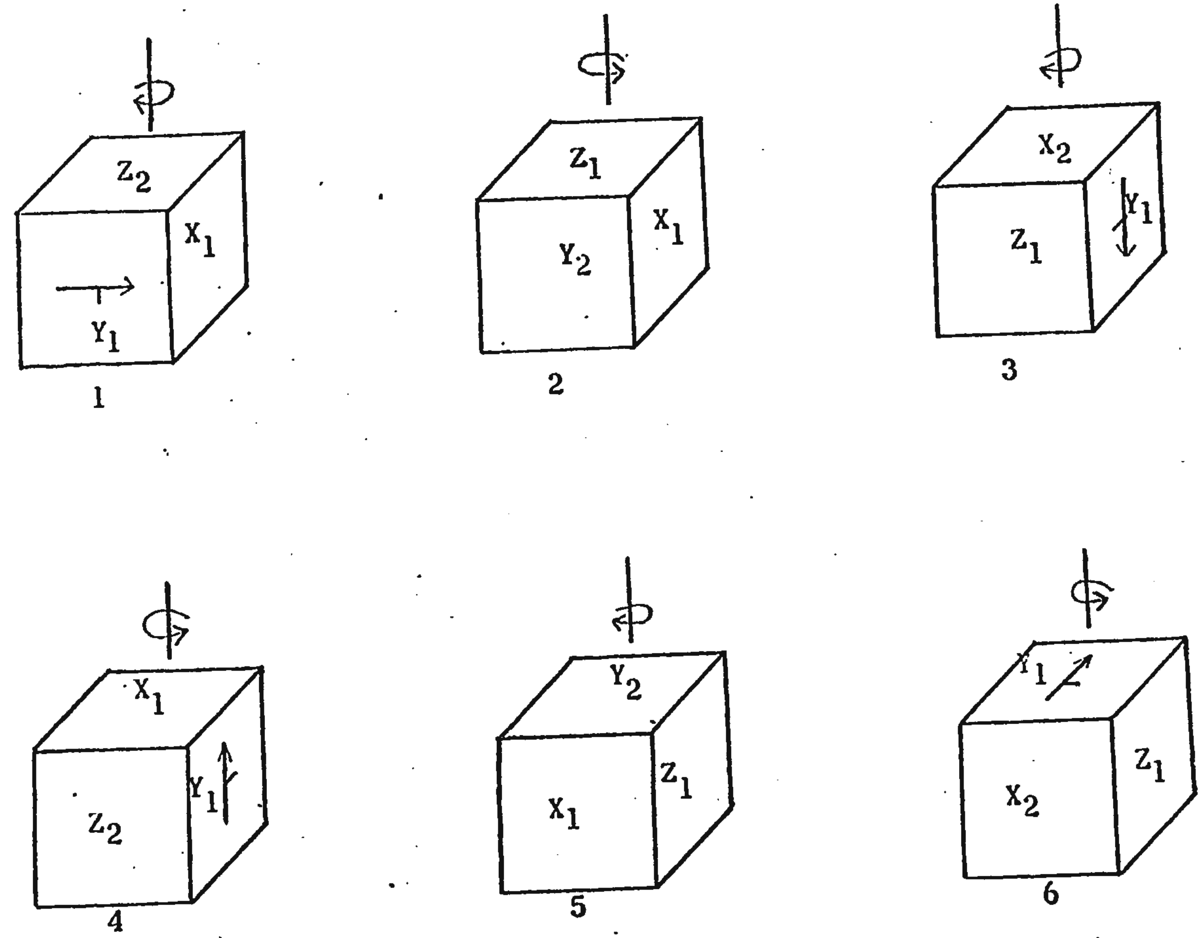


Fig. 15 POSITIONS OF THE CUBICAL SPECIMENS FOR THE MEASUREMENT OF AZIMUTH AND DIP (SEE TEXT)

From columns (1) and (2), the ratio $\frac{Y}{X}$ may be calculated as

$$\frac{Y}{X} = \frac{(Y_1)_1 + (Y_1)_2 - (Y_2)_1 - (Y_2)_2}{(X_1)_1 + (X_1)_2 - (X_2)_1 - (X_2)_2} \tag{6.1}$$

Similarly, from columns (3) and (4):

$$\frac{Y}{Z} = \frac{(Y_1)_3 + (Y_1)_4 - (Y_2)_3 - (Y_2)_4}{(Z_1)_3 + (Z_1)_4 - (Z_2)_3 - (Z_2)_4} \tag{6.2}$$

From columns (5) and (6)

$$\frac{Z}{X} = \frac{(Z_1)_5 + (Z_1)_6 - (Z_2)_5 - (Z_2)_6}{(X_1)_5 + (X_1)_6 - (X_2)_5 - (X_2)_6} \tag{6.3}$$

From the above ratios, the values of the components $\frac{R}{X}$, $\frac{R}{Y}$, and $\frac{R}{Z}$ may be obtained as follows:

$$\frac{R}{X} = \left(1 + \frac{Y^2}{X^2} + \frac{Z^2}{X^2}\right)^{1/2} \tag{6.4}$$

$$\frac{R}{Y} = \left(1 + \frac{X^2}{Y^2} + \frac{Z^2}{Y^2}\right)^{1/2} \qquad (6.5)$$

$$\frac{R}{Z} = \left(1 + \frac{X^2}{Z^2} + \frac{Y^2}{Z^2}\right)^{1/2} \qquad (6.6)$$

where

$R = (X^2 + Y^2 + Z^2)^{\frac{1}{2}}$ is the length of the magnetic vector.

From the above, the azimuth, A, and dip, D, may be obtained as follows:

$$A = \tan^{-1} \frac{Y}{X} \qquad (6.7)$$

$$D = \tan^{-1} \frac{Z}{(X^2 + Y^2)^{1/2}} \qquad (6.8)$$

The intensity of magnetization, J, can be obtained by using a similar procedure as in the case of the cylinders discussed in Section 3.4. In the case of cubes, the maximum deflection due to the horizontal component of the magnetic vector in each of the three planes is measured, and its mean value is used instead of δ in equation 3.2.

6.5 Results

An attempt was made to measure the magnetic vectors of all 15 cubes cut from the samples (Section 6.2), but five out of the eight specimens from VR3 turned out to be inhomogeneous, and their magnetization could not be represented by dipole at the cube center. This will be further discussed in Section 6.7.2. Results for the remaining 10 specimens are shown in Table 16 and the sample averages, calculated according to Section 3.5.1 relative to the present horizontal plane, are tabulated in Table 17.

Tilt corrections (Section 3.5.4) were applied to the results in Table 17 on the usual assumption that the beds have been tilted about the strike line by an angle equal to the measured tilt. Most probably this assumption is justified in the case of samples VR2 and VR3, because the observed strike and dip (5^{o}, 29^{o}E) at the 17-fathom shoal from which these samples were taken (Section 6.2) conform to the gross structure of the Virgin Rocks, as reported by Lilly (1965). On the other hand, the assumption may be inapplicable to sample VR1 for the 9-fathom shoal, because the strike and dip in the latter case (113^{o}, 10^{o}S) were probably measured at a local bend in the strata (Deutsch and Lilly, to be published)l In view of this, the "corrected" direction for VR1 could be in error. These directions, relative to both the present and the ancient horizontal plane, are plotted on an equal area net in Figure 16, which shows also the directions of the present acting and axial dipole fields in the Virgin Rocks area.

TABLE 16

REMANENT MAGNETIZATION OF SPECIMENS FROM THE VIRGIN ROCKS RELATIVE TO PRESENT HORIZONTAL PLANE AND TRUE NORTH

SPECIMEN NUMBER	AZIMUTH A (degrees)	DIP D (degrees)	MEAN INTENSITY OF MAGNETIZATION $J \times 10^{-5}$ (emu/cc)
VR1 (1)	222	-35	2.9
(2)	187	32	2.1
VR2 (1)	52	8	0.84
(2)	49	- 6	5.5
(3)	34	34	6.4
(4)	8	-73	3.4
(5)	35	1	1.7
VR3 (1) - (5)	Inhomogeneous magnetization. Results cannot be represented by dipole.		
(6)	52	7	0.58
(7)	340	11	8.9
(8)	116	-32	11

TABLE 17

SAMPLE AVERAGES RELATIVE TO PRESENT HORIZONTAL PLANE AND TRUE NORTH

SAMPLE NUMBER	NUMBER OF SPECIMENS AVERAGED	MEAN AZIMUTH ($\bar{A}$)	MEAN DIP ($\bar{D}$)	MEAN INTENSITY (emu/cc)
VR 1	2	204°	-1°	2.5×10^{-5}
VR 2	5	45°	-13°	3.7×10^{-5}
VR 3	3	46°	-7°	6.8×10^{-5}

Present direction of earth's field at the Virgin Rocks:

$A = 333^\circ$; $D = 70^\circ$

Direction of geocentric axial diploe field:

$A = 0^\circ$; $D = 64\frac{1}{2}^\circ$

TABLE 18

SAMPLE AVERAGES AFTER TILT CORRECTION HAS BEEN APPLIED

SAMPLE No.	AZIMUTH	DIP
VR 1	205°	-11°
VR 2	35°	-26°
VR 3	37°	-23°

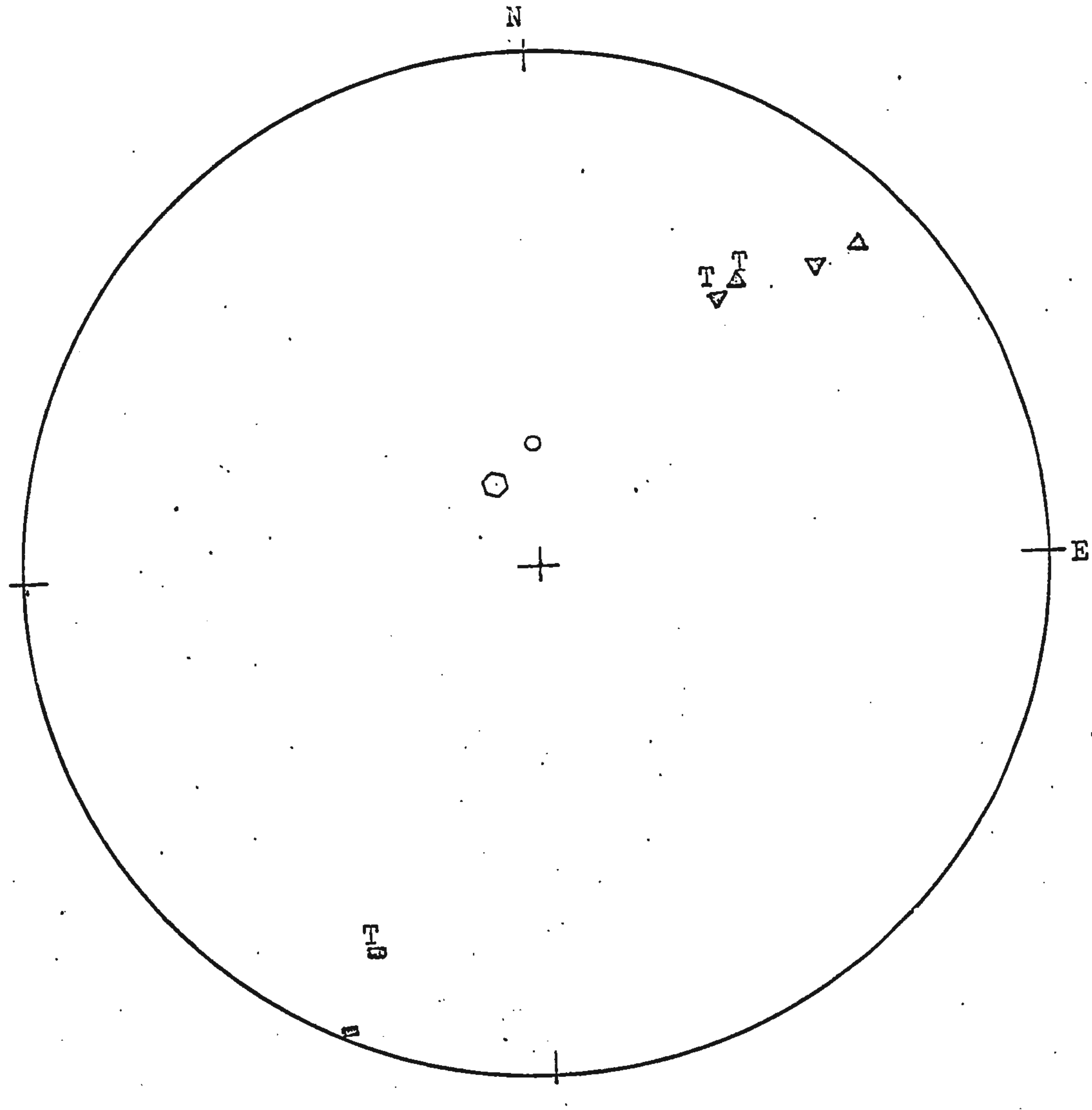


Fig. 16 MEAN DIRECTIONS OF MAGNETIZATION OF SAMPLES FROM THE VIRGIN ROCKS

Solid symbols indicate north poles upward
Hallow symbols indicate north poles downward

△ △T Mean direction for VR 1 before and after tilt correction
▽ ▽T Mean direction for VR 2 before and after tilt correction
▭ ▭T Mean direction for VR 3 before and after tilt correction
○ Theoretical axial dipole field at sampling site
⬡ Direction of the present geomagnetic field at sampling site

6.6 Calculation of the palaeomagnetic pole

From the corrected data, the mean direction of magnetization for the 17-fathom shoal is

$$\bar{A} = 36^\circ; \qquad \bar{D} = -25^\circ$$

where $\bar{A}$ and $\bar{D}$ are the mean azimuth and dip, respectively.

The corresponding palaeomagnetic pole (Section 3.5.3) is at

$$\lambda' = 23^\circ N \qquad \phi' = 91^\circ E$$

where λ' and ϕ' are the present latitude and longitude, respectively of the ancient pole, and the polarity is such that the magnetic north pole falls in the northern Hemisphere.

6.7 Discussion of results

6.7.1 Reliability of results

As mentioned in Section 3.6, a combined orientation and measurement error of 3-5° is acceptable in palaeomagnetism. The results from the Virgin Rocks are not accurate to this level because the method of underwater orientation of samples had not yet been perfected when this initial collection was made, and the orientation marks on the rock surfaces are quoted to be accurate to 5-10° only (Deutsch and Lilly, to be published).

Moreover, the number of samples examined was very small. In proposing certain minimum criteria of reliability in palaeomagnetism, I~ving (1964; p. 102) states: "No result is considered adequate unless it is based on consistent observations from 5 or more separately oriented samples". Clearly, it is not possible to attach much significance to the results from the Virgin Rocks, and no statistical parameters will be quoted.

At the same time, recovery of these samples represents a pioneering application of the methods of underwater geology to palaeomagnetic sampling, and the shoal area of the continental shelf from which they were collected represent new territory for palaeomagnetic investigations. This endows the results obtained here with a special interest, despite the fact that they are inconclusive. Final palaeomagnetic conclusions must wait until the shoals have been studied in detail.

6.7.2 Inhomogeneity of magnetization of samples from the Virgin Rocks

Table 16 shows that the magnetization in 5 out of the 8 specimens from VR3 cannot be represented by a dipole. This may be due to the inhomogeneity of the brecciated rock which contains, along with inclusions of other composition, some granitic pebbles exceeding in volume that of a single cube. Although care was taken to cut the cubes from the apparently least inhomogeneous region of each sample, this could not ensure that the distribution of the relevent ferromagnetic minerals would be always symmetrical with respect to the cube axes. The inhomogeneous nature of VR3 was evident also from wide variations in its intensity of magnetization, J: e.g., two of the cubes from sample VR3 which did yield dipole directions and were originally about 15cms. apart in the rock, have intensities of 11 x 10^{-5} and 0.6 x 10^{-5} e.m.u./cm^3, respectively; in the ratio 18:1.

6.7.3 Magnetization directions and stability

While the palaeomagnetic directions are inconclusive for the reasons stated above, Table 16 shows that the remanent vectors in nine out of ten specimens have low magnetic dips ranging from $+35^{\circ}$ to -35°; hence they make large angles with the present direction of the geomagnetic field at the Virgin Rocks locality and also with that of an axial geocentric dipole (see Fig. 16). This indicates at least partial stability.

The NRM of specimen VR3 (8) was remeasured after about 5 months' storage in the laboratory, with little change in the azimuth and dip. The specimen was then subjected to stepwise alternating field demagnetization in peak fields up to 324 oe. Treatment at 27 oe. caused no significant change in the direction of magnetization, while a 54 oe. field resulted in a change of the order of 10°. Subsequent storage in the earth's field, however, revealed that a VRM with low time-constant had been acquired, as the magnetization vector was then significantly deflected by the earth's field in the course of a few hours. Further treatment in a peak field of 80 oe. failed to remove this viscous component, but subsequent demagnetization at 150 and 250 oe. caused the magnetization once more to assume directions with low dip, though appreciably different from that of the NRM before demagnetization. Finally, continuation of the treatment at 324 oe. resulted in a further deflection of the magnetization vector by 30° or so, suggesting that part of the VRM remained.

The above tests confirm that the NRM before demagnetizaton was at least partially stable. A more conclusive result should be obtainable from a combination of thermal and alternating field demagnetization of these rocks; e.g. As and Zijderveld (1958) found that treatment in a maximum a.c. field of 300 oe. combined with thermal demagnetization up to 150°C was sufficient to remove unstable components in the rock samples they studied.

After tilt correction, all samples have low upward dips (Table 18), but while the directions of VR2 and VR3 agree very closely with one another, their azimuths are quite different from that of VR1. This discrepancy may be related to the fact that the strike and dip measured at the 9-fathom shoal were anomalous (Section 6.2).

Pending a complete palaeomagnetic study of the Virgin Rocks, it is of interest to compare briefly the ancient pole position (23°N, 91°E) that would correspond to the mean direction of magnetization of the 17-fathom shoal (Section 6.6) with that inferred by Nairn et al (1959) from late Precambrian rocks on the Avalon Peninsula (11 N, 122 W). While both poles are in low latitudes, the co-ordinates of the latter depart rather widely from those inferred at the shoal or even from the pole position antipodal to them (23°S, 89°W).

Despite their inconclusive nature, these results are somewhat encouraging, because they suggest that the magnetization of the rocks is at least partially stable, while greatly diverging in direction from that of the present earth field. This, together with the fact that the recovery of oriented bedrock samples from high shoals far from land had been demonstrated, should justify confidence in the results of greatly amplified palaeomagnetic studies on the Grand Banks.

SUMMARY AND CONCLUSIONS

Instrumentation

The design of an astatic magnetometer suitable for the measurement of weak remanent magnetization in rock specimens has been discussed. Along with high sensitivity, a practical instrument should have a relatively low period T and a high P/I ratio, where P is the magnetic moment of either magnet in the astatic system and I the total moment of inertia of the system. Sensitivity, T and P/I are not independent quantities, and the actual design must involve some compromise.

In the present case, P/I was made large through choice of magnets composed of a material ('Magnadur 3') having large specific dipole moment while being magnetized transversely to their long axes. This means that P is large, while I can be reduced drastically because of the vertical mounting of such magnets. The inertia was further reduced by careful design of the frame and mirror of the astatic system. 'Magnadur 3' proved superior to such materials as Alcomax IV and Ferroxdure traditionally used in astatic systems (e.g. Blackett, 1952; Roy, 1963), by allowing the magnets to attain larger specific dipole moments in the transverse direction; moreover its unusually large coercive force (H_c = 3000 oe.) tends to give them great stability in the presence of demagnetizing fields. In the completed instrument P/I was found to be 1.35 x 10^3 c.g.s.

units.

An astaticism of 6000 or more was achieved with the aid of small magnetized particles of "Magnadur 3" which were placed directly upon the magnet surfaces in appropriate positions. This removed much of the awkwardness involved in the use of "trimming magnets" apart from the advantage of using material of high coercive force.

After construction of the magnetometer, including three orthogonal sets of Helmholtz coils for nulling the earth's field, the instrument was installed in a laboratory in the second floor of the Science Building where permanent magnetic field gradients and disturbances from time-varying d.c. fields appeared to be at a minimum. Even so, the location was not sufficiently quiet magnetically to have made it practical to aim for the highest sensitivity. Using a gold ribbon suspension of torsion constant $\sigma = 1.41 \times 10^{-2}$ dyne-cm/rad, the actual reciprocal sensitivity was 3.9×10^{-7} oe. per mm deflection at 1.8 m. distance to the scale with T = 4.6 sec. However, this permitted the remanent magnetization of specimens with minimum intensity $J = 1 \times 10^{-5}$ e.m.u./cm to be measured quickly and accurately.

When required, one can easily increase the sensitivity without changing the design of the astatic system, by substituting a thinner or longer suspension fibre, or by increasing the magnetometer-scale distance. Because of the large P/I value even a tenfold increase in sensitivity, permitting measurement of the more weakly magnetic sedimentary rocks, can be accomplished with relatively short period ($T < 15$ sec.).

After the magnetometer had been installed it was discovered that short-period mechanical vibrations transmitted through the weak floor of the laboratory caused the light spot at the scale to oscillate with several mm. amplitude; this made it impossible to operate the instrument. The magnetometer was therefore fitted with a specially designed, top-heavy spring suspension to lengthen the natural period. This modification succeeded in virtually eliminating all vibrations of the light spot.

Palaeomagnetic investigations

Oriented rock samples were collected from various parts of Newfoundland for the palaeomagnetic investigations.

Remanent magnetization directions and intensities of 28 basalt samples from Henley Harbour and Table Head on the Labrador coast were measured with the astatic magnetometer, and a Fisher analysis was performed. The mean direction of NRM was $\bar{A} = 21.2^{\circ}$, $\bar{\bar{D}} = 76.5^{\circ}$ with $K = 15.0$ and $\alpha_{95} = 7.6^{\circ}$. This corresponds to a position of the geomagnetic pole at the present co-ordinates: 74°N, 22°W. This is too far from the polar wandering curve relative to North America and suggests that the magnetization vectors in these rocks may have been pulled towards alignment with the earth field long after deposition of the basalt. The presence of an unstable component is also suggested by the relatively small angle between the mean magnetization direction and that of the present earth's field or axial dipole field.

On the other hand, alternating field demagnetization tests carried out on some specimens showed the NRM to be mostly stable, while thin sections from two basalt specimens revealed that hematite, probably of low temperature, secondary origin, was much more prominent than magnetite. Pending confirmation of these results by further studies, it was concluded that the observed NRM could contain a prominent hard component of post-depositional origin, for the destruction of which, and the unmasking of any primary TRM residing in the magnetite, thermal demagnetization may be required.

The NRM of three samples of sedimentary rock (arkose) underlying the basalt, has been measured. The directions (reported in Table 6A) are very similar to those of the overlying basalts, but no significant conclusions can be drawn from this result, as the number of samples studied are few.

Preliminary palaeomagnetic investigations of some grey and red argillite samples from three localities in Springdale area of north-central Newfoundland are also reported in this thesis. The rocks are of Ordovician age and relatively unmetamorphosed. The magnetic dips relative to the bedding planes are always low, suggesting that the rocks, at the time of their magnetization, were in low latitudes. Assuming that the beds have not rotated about a vertical axis, the mean magnetization directions at the three sites would correspond to ancient poles at low latitude, but differing widely in longitude both from one another and from the single Ordovician result relative to North America (Collinson and Runcorn, 1960). The NRM of a few specimens, subjected to alternating field demagnetization, proved to be moderately stable in fields up to 324oe, but the study of both the remanent magnetization and of its stability, will have to be greatly enlarged in scope before reliable palaeomagnetic conclusions can be drawn.

Palaeomagnetism results of the samples of microbreccia from the Virgin Rocks shoals of the Grand Banks of Newfoundland are also reported. The investigations appear to be the first reported palaeomagnetic measurements of underwater bed-rock samples, though deep-sea unconsolidated samples have been studied palaeomagnetically. Most of the specimens cut from one of the samples, were found to be too inhomogeneous in their magnetization to permit the NRM to be measured, but the directions obtained in all other cases made significantly large angles with the present earth's field direction. While this indicates partial stability, further work with a much larger collection from the Virgin Rocks is needed to draw any significant conclusions.

<u>Scope for further work</u>

The interesting results obtained from the present study, warrant the continuation of the various projects initiated in this study. The following approach to future investigations may be suggested.

a) <u>Magnetometer</u>: The astatic magnetometer in its present position on the 2nd floor is not conveniently placed, as mentioned before.

In order to permit operation of the instrument at increased sensitivity, it should be placed in a basement room as far as possible from all magnetic materials, d.c. and a.c. sources, etc. The best environment would be in a non-magnetic hut in a relatively isolated location, though this is not essential. Procedures for improving the sensitivity have already been discussed.

b) <u>Remanent magnetization measurements</u>: The NRM measurements reported in this thesis are not fully confirmed for two reasons: (1) Enough samples have not been measured in the case of the Virgin Rocks and Springdale localities, (2) In the case of the Labrador basalts and the Springdale rocks, the closeness of the measured mean direction to the present geomagnetic field direction makes it especially important to establish stability; the presence of a strong viscous component ~~due to the present direction of the earth's field~~ acquired long after the rocks were formed cannot be ruled out.

In the opinion of Clifford (1965), the basalts at Cloud Mountain on the Northern Peninsula of Newfoundland are of the same age as the Labrador Coast basalts, because of the close similarity in the sequence of rocks at both localities, and their association with rocks of known Cambrian age. A study of the NRM and stability of the basalts and sediments at Cloud Mountain, and of a statistically adequate suite of sedimentary rocks underlying the Labrador basalts could lead to significant conclusions on the still undecided problems regarding the age of these rocks and the ancient geography of the region under study.

ACKNOWLEDGEMENTS

I wish to express my thanks to the following people who have helped in the course of the work described in this thesis:

Dr. E. R. Deutsch for his constant encouragement, guidance and help throughout the progress of this study.

Also at the Physics Department of Memorial University, Dr. P. S. Rao, for critically reviewing the manuscript and offering helpful suggestions; and Mr. G. W. Pearce, who carried out alternating field demagnetization tests and also helped the author in the compilation of data with the IBM 1620 computer at the University.

At the Geology Department, Dr. V. S. Papezik, who suggested the collection sites near Springdale and provided other valuable information on the rocks in that area; and Mr. A. C. Nautiyal, who analyzed the thin-sections from Henley Harbour and Table Head.

Dr. E. Irving, formerly of the Dominion Observatory, Ottawa, for suggesting the method of astatizing the magnetometer discussed in Chapter 2.

Dr. A. P. Beavon, Chief Geologist of British-Newfoundland Exploration Co., Ltd., and staff members; in particular, Mr. H. R. Peters, Chief Geologist at the Springdale base camp, arranged for the author's air transport to Labrador and other transportation, and offered much encouragement. Mr. W. Williamson, Springdale camp manager, provided full facilities at the camp.

Mr. J. Helwig of Columbia University, New York, and Mr. D. Batchinsky of the Geological Survey of Canada who suggested to the author some of the collection sites in north-central Newfoundland and also provided valuable geological information in the field.

Mr. A. J. Walsh, Science Assistant in the Physics Department, carried out the photographic work included in this thesis.

I also wish to express my gratitude to Memorial University of Newfoundland for providing financial assistance in the form of a Graduate Fellowship during the years 1964-66.

Financial assistance during the summers of 1965 and 1966 was provided from the National Research Council of Canada Grant A-1946 to Dr. E. R. Deutsch.

The palaeomagnetism of the rocks in the Springdale area was studied with financial help from Grant No. 14-65 by the Geological Survey of Canada to Dr. E. R. Deutsch.

REFERENCES

As, J. A. and Zijderveld, J.D.A. (1958); Magnetic cleaning of rocks in palaeomagnetic research: Geophy. Jour. Royal Astr. Soc., vol. 1, p. 308-319

Black. R. F. (1964); Palaeomagnetic support of the theory of rotation of the western part of the island of Newfoundland: Nature, vol. 202, p. 945-948

Blackett, P.M.S. (1952); A negative experiment relating to magnetism and the earth's rotation: Phil. Trans. Royal Soc. London, ser. A, vol. 245, p. 303-370

Bruckshaw, J. McG. and Robertson, E.I. (1948); The measurement of magnetic properties of rocks: Jour. Sci. Inst., vol. 25, p. 444-446

Bullard, E.C. (1949); The magnetic field within the Earth: Proc. Royal Soc., London, ser. A, vol. 147, p. 433

Chapman, S. and Bartels, J. (1962); Geomagnetism: Oxford, England, Clarendon Press, p. 1049

Chevallier, R. (1925); L8aimantation des laves de l'Etna et l'orientation du champ terrestre en Sicile du XII[e] au XVII[e] siecle; Ann. Phys., 4, p. 5-162

Chikazumi, S. (1964); Physics of magnetism: John Wiley & Sons, Inc., New York, ~~p.~~ 554p

Christie, A. M. (1951); Geology of the southern coast of Labrador from Forteau Bay to Cape Borcubane, Newfoundland: Geological Survey of Canada, Pages 51-13, 19p

Clifford, P. M. (1965); Palaeozoic flood basalts in northern Newfoundland and Labrador: Can. Jour. Earth Sci., vol. 2, p. 183-187

Collinson, D. W. and Creer, K. M. (1960); "Measurements in palaeomagnetism", in Methods and Techniques in Geophysics, vol. 1, ed. S. K. Runcorn, Interscience Publishers, New York, p. 168-210

Collinson, D. W., Creer, K. M., Irving, E. and Runcorn, S. K. (1957); Palaeomagnetic investigations in Great Britain I. The measurement of the permanent magnetization of rocks: Philo Trans. Royal Soc. London, ser. A, vol. 250, p. 71-82

Collinson, D. W. and Runcorn, S. K. (1960); Polar wandering and continental drift: evidence of palaeomagnetic observations in the United States: Geol. Soc. Amer. Bull., vol. 71, p. 915-958

Cox, A. (1964); Angular dispersion due to random magnetization: Geophy. Jour. Royal Astron. Soc. vol. 8, p. 345-355

Cox, A. and Doell, R. R. (1960); Review of paleomagnetism: Geol. Soc. Amer. Bull., vol. 71, p. 645-768

Cox. A., Doell, R. R. and Dalrymple, G. B. (1963); Geomagnetic polarity epochs and Pleistocene geochronometry: Nature, 198, p. 1049-1051

Creer, K. M. Irving, E. and Runcorn, S. K. (1957); Geophysical interpretation of palaeomagnetic directions from Great Britain; Phil. Trans. Royal Soc. London, ser. A. vol. 250, p. 144-156

Deutsch, E. R. and Lilly, H. D; Palaeomagnetic reconnaissance on the continental shelf, 170 km. east of Cape Race, Newfoundland: to be published

Doell, R. R. and Cox, A. (1963); The accuracy of the paleomagnetic method as evaluated from historic Hawaiian lava flows: Jour Geophy. Res., vol. 68, p. 1997-2009

Douglas, G. V. (1953); Notes on localities visited on the Labrador coast in 1946 and 1947: Geol. Surv. of Canada, Paper 53-1

Eade, K. (1962); Geology, Battle Harbour - Cartwright map area, coast of Labrador, Newfoundland: Geol. Surv. of Canada, Map 22-1962

Elsasser, W. M. (1955); Hydromagnetism I. A review: Amer. Jour. Phy., vol. 23, p. 590-609
(1956); Hydromagnetism II. A review: Amer. Jour. Phy., vol. 24, p. 85-110

Fisher, R. A. (1953); Dispersion on a sphere: Proc. Royal Soc. London, ser. A, vol. 217, p. 295-305

Graham, J. W. (1949); The stability and significance of magnetism in sedimentary rocks: Jour. Geophy. Res., vol. 54, p. 131-167

Griffiths, D. H., King, R. F. and Wright, A. E. (1957); Some field and laboratory studies of the depositional remanence of recent sediments: Phil. Mag. Supp. Adv. Phys., vol. 6, p. 306-316

Harrison, C. G. A. (1966); The palaeomagnetism of deep sea sediments: Jour. Geophy. Res. vol. 71, p. 3033-3045

Hospers, J. (1955); Rock magnetism and polar wandering: Jour. Geol., vol. 63, p. 59-74

Howell, L. G. Martinez, J. D. and Statham, E. H. (1958); Some observations on rock magnetism: Geophysics, vol. 23, p. 285-298

Irving, E. (1964); Palaeomagnetism and its application to geological and geophysical problems: John Wiley & Sons, INC., New York, 399p

Irving, E., Stott, P. M. and Ward, M. A. (1961); Demagnetization of igneous rocks by alternating magnetic fields: Phil. Mag., vol. 6, p. 225-241

Johnson, E. A. and McNish, A. G. (1938); An alternating-current apparatus for measuring small magnetic moments: Terr. Mag. Atm. Elect., vol. 43, p. 393-399

Koenigsberger, J. G. (1930); Grossenverhaltnis von remanentem zu induziertem Magnetismus in Gesteinen; Grosse und Richtung des remanenten Magnetismus: Z. f. Geophys. Vol. 6, p. 190-207

Kranck, E. H. (1939); Bedrock geology of the seaboard region of Newfoundland Labrador: Geol. Surv., Newfoundland, Bull. No. 19, 44p

Lilly, H. D. (1965); Submarine examination of the Virgin Rocks area, Grand Banks, Newfoundland: Preliminary note: Geol. Soc. Amer. Bull., vol. 76, p. 131-132

Maclean, H. J. (1947); Geology and Mineral Deposits of the Little Bay Area: Geol. Surv. Nfld., Bull. 22

Mullard and Co. (1964); Catalogue on Permanent Magnets

Nagata, T. (1953); Rock magnetism: Maruzen and Co., Tokyo, 225p

Nagata, T. (1961); Rock magnetism: Revised edition, Maruzen and Co., Tokyo, 350p

Nairn, A.E.M., Frost, D.V., and Light, B.G. (1959); Palaeomagnetism of certain rocks from Newfoundland: Nature, vol. 183, p. 596-597

Neél, L. (1951); L'inversion de l'aimantation permanente des roches: Ann. Geophys., vol. 7, p. 90-102

Neél, L. (1955); Some theoretical aspects of rock magnetism: Phil. Mag. Supp. Adv. Phys., vol. 4, p. 191-243

Papezik, V. S. (1963); Unpublished geological report on Nicky's Nose and Rushy Pond Prospects to British Newfoundland Exploration Ltd. 4p

Roy, J. L. (1963); The measurement of magnetic properties of rock specimens: Domin. Observ., Ottawa, vol. 27, no. 9, p. 419-439

Schuchert, C. and Dunbar, C. O. (1934); Stratigraphy of western Newfoundland: Geol. Soc. Amer. Mem. No. 1

Stacey, F. D. (1963); The physical theory of rock magnetism: Adv. Phys., vol. 12, p. 46-133

Thellier, E. (1938); Thesis, University of Paris

Thellier, E. and Rimbert, F. (1954); Sur l'analyse d'aimantations fossiles par action de champs magnétiques alternatifs: C.R. Acad. Sci., Paris, vol. 239, p. 1399-1401

Uyeda, S. (1958); Thermo-remanent magnetism as a medium of palaeomagnetism, with special reference to reverse thermo-remanent magnetism; Japan. Jour. Geophys., vol. 2, p. 1-123

Wanless, R. K., Stevens, R. D., Lachance, G. R., and Rimsaite, R.Y.H. (1965); Age determinations and geological studies. Part 1 - Isotopic ages, Report 5: Geol. Surv. of Canada, Paper 64-17 (Part 1) 126p

Watkins, N. D. (1961); Studies in palaeomagnetism: Unpublished M.Sc thesis, University of Alberta, Edmonton, Alberta

Watson, G. S. and Irving, E. (1957); Statistical methods in rock magnetism: Mon. Not. Royal Astr. Soc. Geophys. Supp., vol. 7, p. 289-300