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A Palaeointensity Method for Use with Highly Oxidised Basalts, and Application to Some Permian Volcanics

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Abstract. A palaeointensity technique is proposed for use with basalts that have undergone high temperature oxidation and whose magnetic minerals are in their most advanced oxidation state. In such basalts the main magnetic mineral is haematite, although there is usually a minor constituent of magnetite present. Because magnetite is a very much stronger magnetic mineral, typically the total remanence of these basalts is derived about equally from each mineral. Such a two-mineral assemblage is quite unsuitable for use in the Thelliers' method. However they are well distinguished by their coercive force spectrum. Therefore a modification to Shaw's method enables palaeo-intensities to be determined separately for each mineral. The low alternating field demagnetization spectrum (≤ 150 mT) provides the palaeointensity for the magnetite fraction and high field alternating field demagnetization (≥ 500 mT) the palaeointensity from the haematite fraction. Agreement between these values provides a powerful consistency check.

Application of this technique to samples of the Permian Exeter traps (U.K.) and the Esterel volcanics (France) shows that it can be used with success. The Exeter traps yield a mean VDM of $6.11 \pm 1.12 \times 10^{22}$ Am² and the Esterel volcanics a mean of $4.03 \pm 1.43 \times 10^{22}$ Am². These values are higher than have previously been suggested for the Permian dipole moment. Standard deviations of the VDMs are 42.9% and 40.6% respectively. This suggests that fluctuations in dipole intensity are the main source of scatter as has been observed in VDMs for the past 5 million years.

Key words: Geomagnetic palaeointensity – Permian volcanics – Two-component remanence.

Introduction

It is well known that the determination of palaeointensity is much more difficult and much less reliable than that of palaeodirection. So that, apart from work in archaeomagnetism, there has been little effort in determining the variation of the magnitude of the earth's dipole moment on the geological time-scale. The main

reasons for this are that the procedures are usually very time-consuming and even then it is not guaranteed that a reliable result will be forthcoming. As a result many attempts have been made to circumvent the classical Thelliers' method (Thellier and Thellier, 1959) with varying degrees of success. The main problem is that heating samples to a temperature above the Curie point (total TRM methods) or successive heatings (Thelliers' method) may cause alteration to the magnetic mineralogy. In Thelliers' method the alteration can be monitored and use made of results prior to the alteration occurring. In total TRM methods, such as Shaw (1974), comparison of coercive force spectra before and after heating is used to monitor whether or not alteration has occurred. We have extended Shaw's method for use in special cases so that an additional consistency check is obtained.

In this paper we consider only the problem of determining palaeointensities from basaltic rocks. We propose a method that makes use only of highly oxidised basalts (Class 5 of Watkins and Haggerty, 1967 or Class 5 and 6 of Haggerty, 1976). The high temperature oxidation in these basalts took place during initial cooling and at temperatures above the Curie temperatures of magnetite and haematite. There are two advantages in using rocks in this oxidation state. First, as the magnetic minerals are in their most advanced oxidation state, further heating should not cause further oxidation. Second, the main magnetic mineral in these basalts is haematite, but there will always be a minor constituent of magnetite present. Because magnetite has saturation magnetization about 200 times that of haematite, its presence in a ratio $\sim 1/200$ times that of haematite will contribute equal magnetizations from the two minerals to the total TRM. Furthermore magnetite and haematite have very different coercive force spectra. In special cases the coercive force of magnetite may extend up to about 0.3 T (Evans and McElhinny, 1969) whereas that of haematite probably extends up to 2.5 T (Rimbert, 1959). Thus it may be possible to determine separate palaeointensities for the low coercivity region (from magnetite) and the high coercivity region (from haematite) using Shaw's method. This would serve as an additional consistency check. Agreement between palaeointensities determined for each mineral fraction would lend much weight to the results obtained.

The New Method

Background

Fundamental to any palaeointensity technique is the relationship (Nagata, 1943) relating the NRM (J_n) of a rock sample with the TRM (J_l) induced in a known magnetic field.

$$\frac{J_n}{J_l} = \frac{F_a}{F_l} \quad (1)$$

where F_a and F_l are the palaeointensity and the laboratory field respectively. The application of this relationship to palaeointensity determinations is hampered by:

- (1) the acquisition of secondary components of magnetization since formation of the rock;
- (2) the decay of the primary components of magnetization; and
- (3) chemical changes that occur during the laboratory heating.

To overcome the first two problems, alternating field (AF) demagnetization has been used to isolate the more stable components of NRM (Van Zijl et al., 1962; Smith, 1967a; McElhinny and Evans, 1968; Doell and Smith, 1969). Sometimes thermal demagnetization (Smith, 1967a) is used, but this then reduces to a form of Thelliers' method, because there will be no consistency between successive temperature intervals until the secondary components are eliminated. The comparison of successive temperature intervals or coercive force intervals then modifies Eq. (1) to:

$$\frac{J_n(T_1, T_2)}{J_l(T_1, T_2)} = \frac{F_a}{F_l} \quad (2)$$

or

$$\frac{J_n(\tilde{H}_1, \tilde{H}_2)}{J_l(\tilde{H}_1, \tilde{H}_2)} = \frac{F_a}{F_l}$$

where (T_1, T_2) and $(\tilde{H}_1, \tilde{H}_2)$ describe the blocking temperature interval and range of alternating magnetic field respectively.

Although the Thelliers' method will detect the point at which chemical changes occur on heating, the alternating field methods require heating to the Curie temperature and chemical changes can only be detected by comparison of the coercive force spectrum before and after heating. Various attempts to minimize chemical alteration have been attempted. Tanguy (1975) tried rapid heating in a preheated furnace, Khodair and Coe (1975) used a vacuum and Kono and Tanaka (1977) used a reducing atmosphere of nitrogen or argon gas. None of these has been particularly successful.

The Thelliers' method (Thellier and Thellier, 1959) and modified versions (Coe, 1967a; Kono and Ueno, 1977; Domen, 1977) has been applied to young basaltic rocks and palaeointensities have been determined using the low to intermediate temperature ranges (e.g. Coe, 1967a; Coe and Grommé, 1973; Coe et al., 1978; Champion, 1980). The problem is the long time required for step-wise heatings and the interpretation of

non-ideal behaviour of the NRM-PTRM relationships or Arai diagrams (Arai, 1963; Nagata et al., 1963) as has been discussed by Coe (1967b) and Levi (1975). Also the Thelliers' method is unsuitable for rocks possessing thermally discrete components (i.e. the blocking temperatures are not well distributed) where all the magnetization is acquired on cooling through a relatively narrow temperature range (McElhinny and Evans, 1968).

The method described by Shaw (1974) compares coercive force spectra of NRM and TRM and also compares the coercive force spectra of ARM before and after a single heating to the Curie temperature to demonstrate the absence of alteration or to identify a coercivity region which has not been affected by this alteration. Kono (1977, 1978) applied Shaw's method to volcanic rocks up to 34 m.y. in age and observed that only a few samples satisfied the 'ideal' conditions set out by Shaw (1974). He concluded that a reliable palaeointensity could be determined, even though the ideal condition was not satisfied, by introducing an appropriate correction term.

We now extend Shaw's method to highly oxidized basalts containing magnetite and haematite with different coercive force spectra.

Procedure

- (1) The NRM of the specimen is progressively demagnetized in alternating magnetic fields up to some maximum peak field \tilde{H}_m . The demagnetized NRM is measured at each step.
- (2) An ARM (ARM1) is given to the specimen by applying an alternating field (\tilde{H}_m) whose amplitude is reduced to zero in the presence of a steady magnetic field of 0.05 mT. The ARM1 is then progressively demagnetized as in (1).
- (3) An IRM (IRM1) is given to the specimen in increasing steps up to fields of about 2.2 T. The magnetization is measured after each step.
- (4) The specimen is given a TRM in a field of about 0.05 mT and it then progressively demagnetized as in (1).
- (5) An ARM (ARM2) is given to the specimen as in (2) and then progressively demagnetized as in (1).
- (6) An IRM (IRM2) is given to the specimen in increasing steps as in (3).

Low Coercivity Spectrum

A palaeointensity value for the magnetite fraction may be determined as in Shaw's method using only the low coercivity region ($0 \leq \tilde{H} \leq 150$ mT) of NRM, TRM and ARM. A comparison is then made of NRM and TRM and ARM1 and ARM2 in this range of demagnetizing fields. The following conditions need to be satisfied to obtain a palaeointensity value.

(a) The magnetite fraction must be sufficient to produce a significant contribution to the total NRM.

(b) At least a part of the low coercivity region must be unaffected by secondary magnetic components or the decay of NRM.

At any alternating field \tilde{H} up to the maximum available \tilde{H}_m , the magnetization will be made up of two components, one due to magnetite and the other due to haematite, so that

$$J_n(\tilde{H}) = J_{mn}(\tilde{H}) + J_{hn}(\tilde{H}), \quad (3)$$

$$J_t(\tilde{H}) = J_{mt}(\tilde{H}) + J_{ht}(\tilde{H}) \quad (4)$$

where $J_{mn}(\tilde{H})$ and $J_{mt}(\tilde{H})$ are the contributions from magnetite and $J_{hn}(\tilde{H})$ and $J_{ht}(\tilde{H})$ are the contributions from haematite. From Eq. (2) then

$$\frac{J_{mn}(\tilde{H}) + J_{hn}(\tilde{H})}{J_{mt}(\tilde{H}) + J_{ht}(\tilde{H})} = \frac{F_a}{F_t}. \quad (5)$$

We now assume that in the range $0 \leq \tilde{H} \leq 100$ mT the contribution to the total coercive force spectrum from haematite is very much less than that from magnetite. Then $J_{hn}(\tilde{H})$ and $J_{ht}(\tilde{H})$ may be considered as constants and

$$J_{mn}(\tilde{H}) = \frac{F_a}{F_t} J_{mt}(\tilde{H}) + \text{constant}; \quad 0 \leq \tilde{H} \leq 100 \text{ mT}. \quad (6)$$

This expression describes a linear relationship between NRM and TRM with slope F_a/F_t providing that the magnetic mineralogy in this coercivity region has not been altered. These alterations may be investigated by comparing ARM1 with ARM2 in the range $0 \leq \tilde{H} \leq 100$ mT (Shaw, 1974; Kono, 1977; 1978). If it is assumed that the change in TRM (or NRM) capacity is the same as the change in ARM capacity, then the gradient of the best fitting line of ARM1 against ARM2 may be used as a correction term for alterations (Kono, 1978). If the NRM-TRM and ARM1-ARM2 plots show linear relationships with slopes m_1 and m_2 respectively in the coercivity range corresponding to $0 \leq \tilde{H} \leq 100$ mT, then

$$F_a(m) = \frac{m_1}{m_2} F_t. \quad (7)$$

High Coercivity Spectrum

Suppose the coercivity spectrum of magnetite is completely eliminated at \tilde{H}_e , where $0 \leq \tilde{H}_e \leq \tilde{H}_m$. Then at any alternating field \tilde{H} in the range $\tilde{H}_e \leq \tilde{H} \leq \tilde{H}_m$ only the haematite component remains and Eqs. (3) and (4) become:

$$\frac{J_n(\tilde{H})}{J_t(\tilde{H})} = \frac{J_{hn}(\tilde{H})}{J_{ht}(\tilde{H})} \quad \tilde{H}_e \leq \tilde{H} \leq \tilde{H}_m. \quad (8)$$

Haematite becomes saturated in fields of about 2.5 T, so it is not a practical proposition to attempt the comparison of ARMs, since alternating fields of this magnitude are difficult to produce. To detect changes in the haematite fraction before and after heating an IRM technique has therefore been used (Carmichael, 1967). For the present study a maximum D.C. field of 2.2 T was available and this is sufficiently close to that required to saturate haematite. As the final step before heating and after heating (steps 3 and 6), the IRM acquired by the specimen is measured in increasing steps.

Suppose the magnetite grains are saturated at some steady D.C. field H_0 ($0 \leq H_0 \leq 2.2$ T). Then the IRM spectrum for fields greater than H_0 is due entirely to haematite. At any field H in this range ($H_0 \leq H \leq 2.2$ T) the IRMs before and after heating can be written as:

$$\begin{aligned} I_1(H) &= M_1 + h_1 \\ I_2(H) &= M_2 + h_2 \end{aligned} \quad H_0 \leq H \leq 2.2 \text{ T} \quad (9)$$

where M_1 and M_2 are the saturation IRMs of magnetite, h_1 and h_2 are the IRMs of haematite before and after heating (subscripts 1 and 2 respectively).

If the magnetization of haematite has been altered by a factor k_2 after heating ($k_2 \sim h_1/h_2$)

$$\frac{I_1(H) - M_1}{I_2(H) - M_2} = k_2 \quad H_0 \leq H \leq 2.2 \text{ T}$$

and

$$I_1(H) = k_2 I_2(H) + (M_1 - k_2 M_2). \quad (10)$$

This represents a straight line with slope k_2 which may be determined from a plot of IRM1 versus IRM2. This is analogous to the ARM1-ARM2 plot in the low coercivity range. If the NRM-TRM relationship is linear with slope k_1 from Eq. (8) in the range $\tilde{H}_e \leq \tilde{H} \leq \tilde{H}_m$, then from Eq. (1)

$$F_a(h) = \frac{k_1}{k_2} F_t. \quad (11)$$

The complete procedure is illustrated in Fig. 1. The overall requirement is that $F_a(m) = F_a(h)$. So that even though the method has to rely on corrections for any alterations that occur, their justification ultimately is that the two independently determined values of the palaeointensity agree. This adds an extra dimension to the point made by Kono (1978) that such "corrected" NRM-TRM slopes may still give correct palaeointensities in certain cases. Senanayake et al. (1982) have demonstrated that, for oxidised rocks, the correction factor proposed by Kono (1978) always produces palaeointensity values in agreement with that determined by the Thellier's method.

High Field A.F. Demagnetization Apparatus

Because of the requirement that high alternating fields be used to obtain at least part of the coercivity spectrum of haematite beyond that of magnetite, a new high field device has had to be constructed. Basically it is a very much larger version of that described by McElhinny (1966) and currently in use in the laboratory at the Australian National University (A.N.U.).

The solenoid is wound with bicalex covered rectangular cross-section copper wire of size 3.14 mm \times 1.26 mm. The bicalex cover adds a further 0.11 mm to each dimension, and the cross-section is equivalent to 13 SWG, making it possible to pass about 20 A through it without any special cooling arrangement. The solenoid has 6940 turns, has internal diameter 9 cm with 78 turns per layer for 89 layers, giving a length of 25.7 cm and external diameter of 34.7 cm. A thin sheet of mylar (thickness 0.008 cm), capable of with-

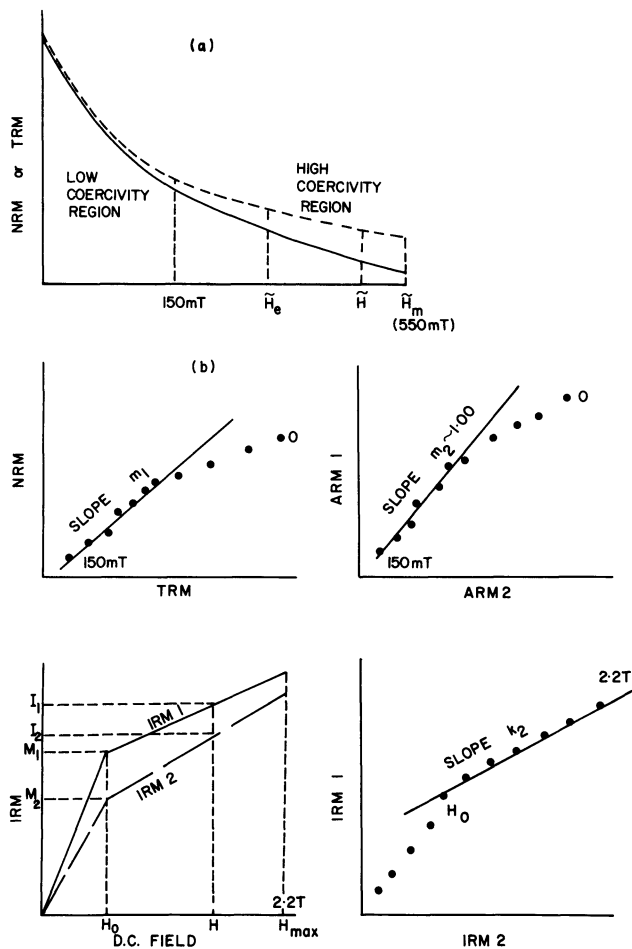


Fig. 1a-c. Illustration of the new palaeointensity technique. **a** Comparison of A.F. demagnetization of NRM and TRM. The magnetite fraction predominates in the low coercivity region in fields < 150 mT. The haematite fraction predominates in the high coercivity region above some field H_e representing the highest coercivity of magnetite; **b** The low coercivity palaeointensity compares NRM and TRM up to fields of 150 mT, and ARM1 and ARM2 before and after heating to the Curie temperature. This corresponds to Shaw's (1974) method as extended by Senanayake and McElhinny (1982); **c** Comparison of the acquisition of IRM before (IRM1) and after (IRM2) heating. Above H_0 the magnetite grains are saturated. The slope k_2 of the IRM1-IRM2 plot at fields $> H_0$ determines the factor by which the magnetization of haematite has been altered on heating

standing a voltage difference of about 5,000 V was glued between each layer with Araldite 'D', a good electrical insulator. The solenoid weighs about 160 kg, has a D.C. resistance of 21.5 ohms and produces a peak alternating field of 36.6 mT/A (rms) as determined by the output of a small search coil placed at the centre of the solenoid.

The demagnetizing solenoid is connected in series with a bank of capacitors (60 kV oil filled) of 2.65 μ F to form a series resonance circuit at 50 Hz. The tuned circuit has a Q factor of about 40 and the apparatus produces peak fields up to 0.55 T under normal conditions with a supply voltage of 485 V (rms). A further 17% increase in the maximum peak field can be obtained using the overvoltage connections on the supply variac. Smooth raising and lowering of the current in

the solenoid is achieved using a motor driven variac which is supplied from second variac connected between two phases of the normal three phase supply giving a source of 415 volts. Because of dangers produced by the high voltage magnification across the solenoid and the capacitors, they are all located behind earthed wire screens.

During trial runs some A.F. demagnetization experiments carried out using the system showed that some spurious magnetization was induced in specimens. Investigation showed that this was not an ARM or RRM but was related to the automatic switching off of the A.C. power through the circuit. The switching gave rise to a very short term high amplitude uni-directional spike through the solenoid. The effect of this would be similar to a lightning effect producing an IRM component. It is apparently due to the capacitors discharging some accumulated charge through the solenoid. To overcome this the A.C. power through the solenoid is never switched off but maintained near zero. The specimen is removed from the coil with the tumbling mechanism still operating.

Application to Some Permian Volcanics

Sample Descriptions

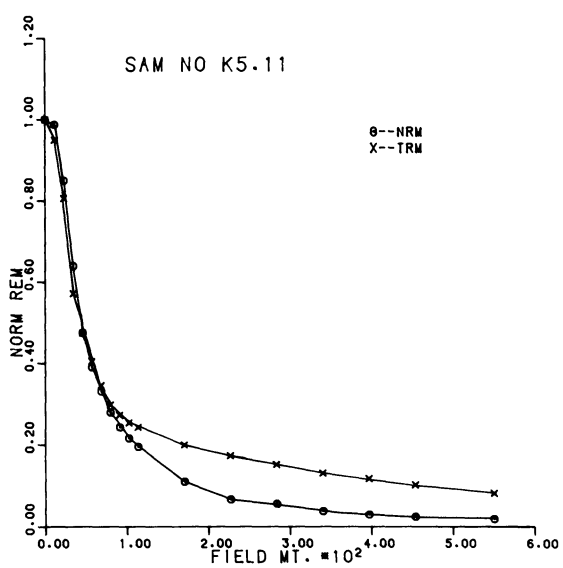
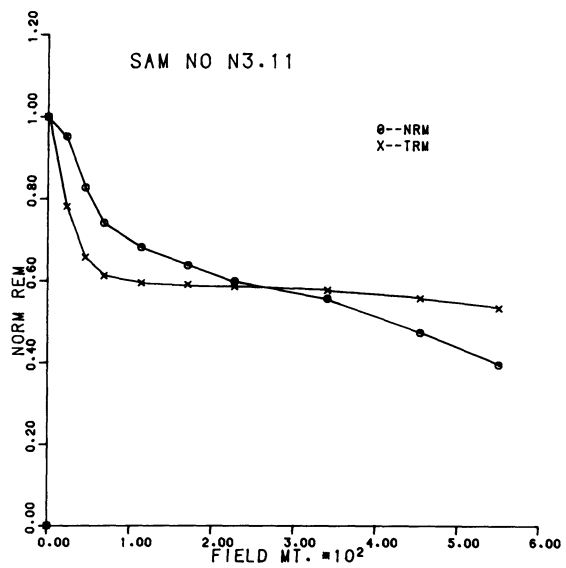
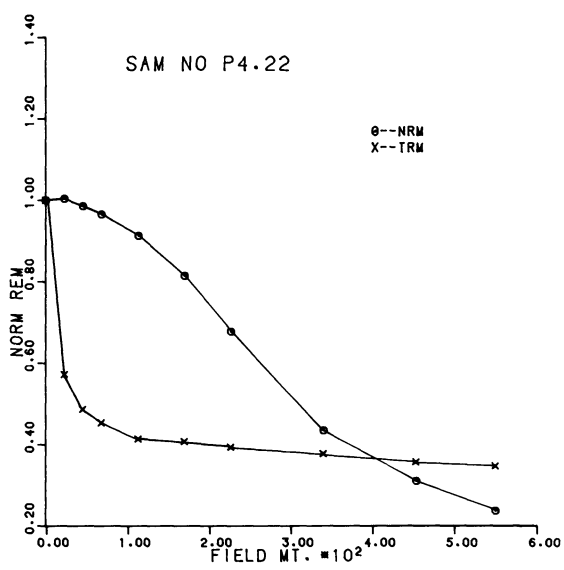
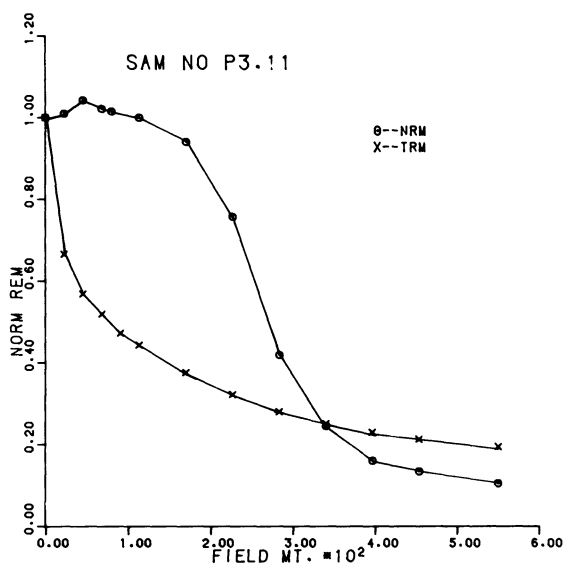
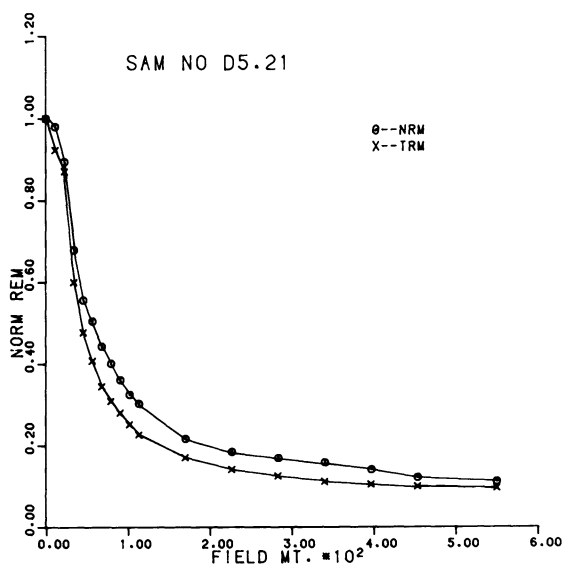
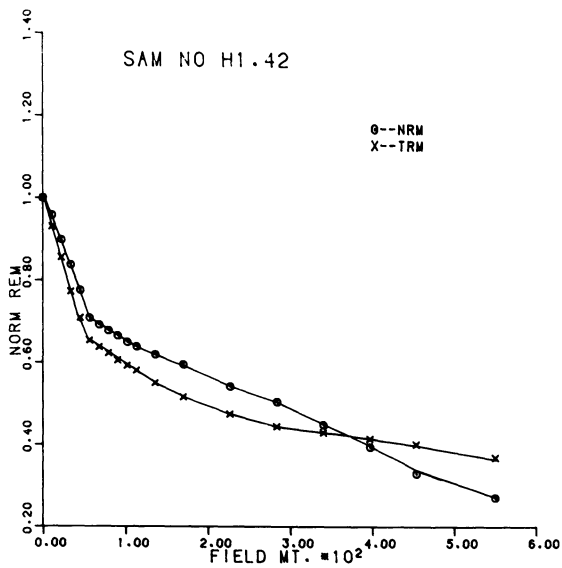
Some very oxidized basalts have been collected from the Exeter traps in England and the Esterel volcanics in southern France, both of Permian age. The A.F. demagnetization characteristics of many of these volcanics described by Zijdeveld (1967; 1975) made it clear that, because they were so resistant to high alternating fields, they were of the type required for the palaeointensity technique.

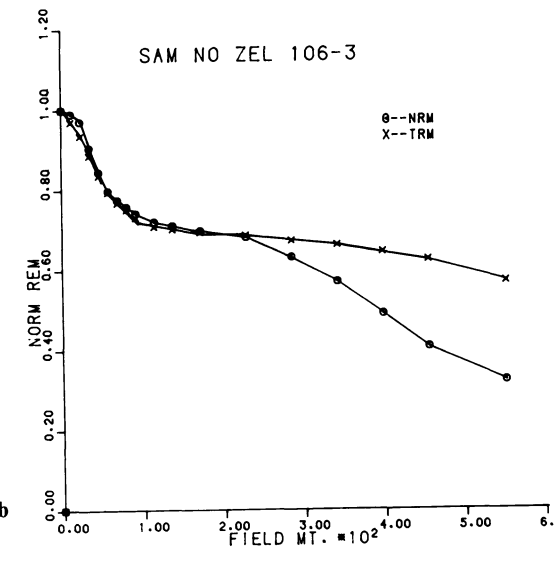
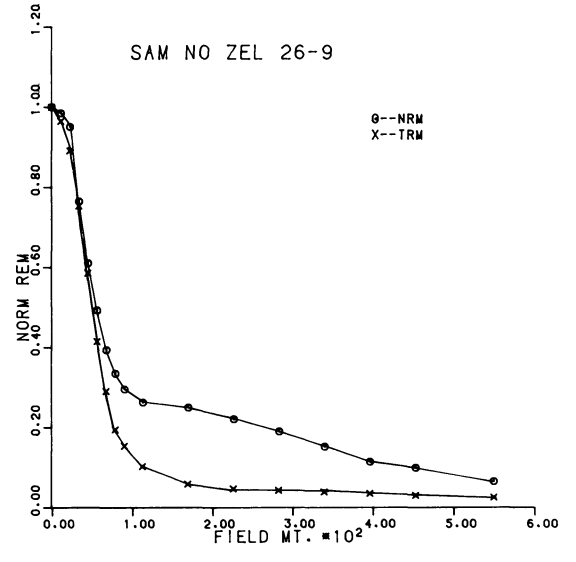
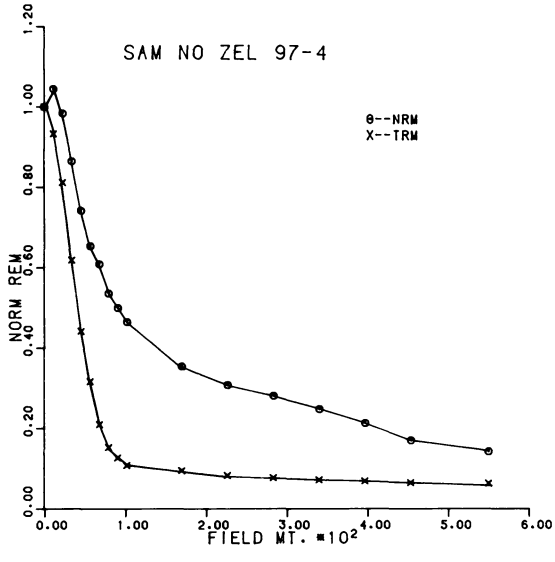
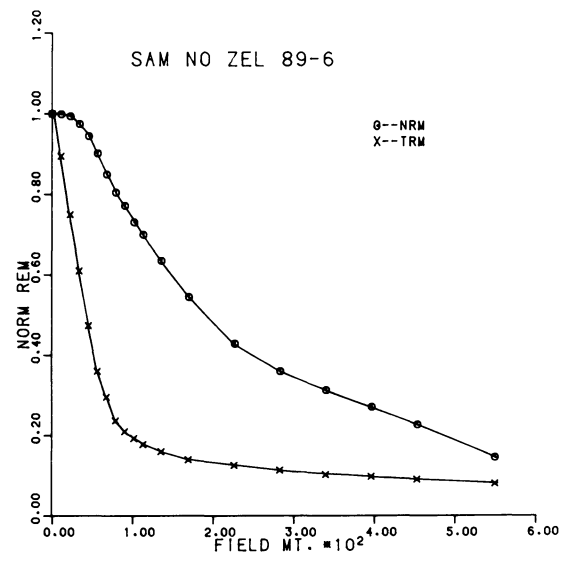
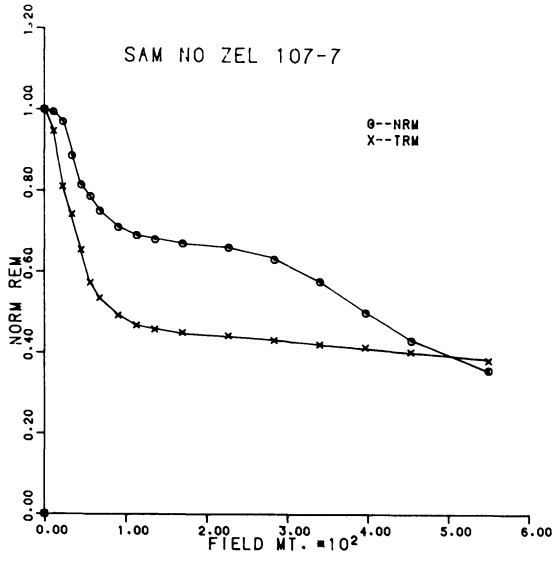
Previous work on the Exeter traps has been carried out by Creer (1957) and Zijdeveld (1967), and descriptions of the localities and geology are adequately described by them. A K-Ar age of 279 ± 6 m.y. has been determined by Miller et al. (1962). The detailed A.F. demagnetization experiments of Zijdeveld (1967) showed that there were 'soft' (low coercivity) and 'hard' (high coercivity) components present in these rocks, but that these had the same palaeodirection, presumed to be primary. This clearly suggests the two mineral magnetite-haematite system required for the palaeointensity method. Zijdeveld (1967) reports a mean palaeodirection of $D = 198^\circ$, $I = -25^\circ$ for these rocks. Twenty-four unoriented blocks were collected from quarries in the Exeter traps at Dunchideock (D), Killerton (K), Heazville (H), Duns Moor (N) and Pocombe (P).

Zijdeveld (1975) has carried out very extensive work on the Esterel volcanics of southern France. Descriptions of the geology and sampling localities have been given by him in some detail. The volcanics are of Late Permian age. Nine specimens from six sites having the most suitable magnetic characteristics were kindly provided by Dr. Zijdeveld from his collection. All these specimens had not been treated either thermally or by alternating fields.

Experimental Results

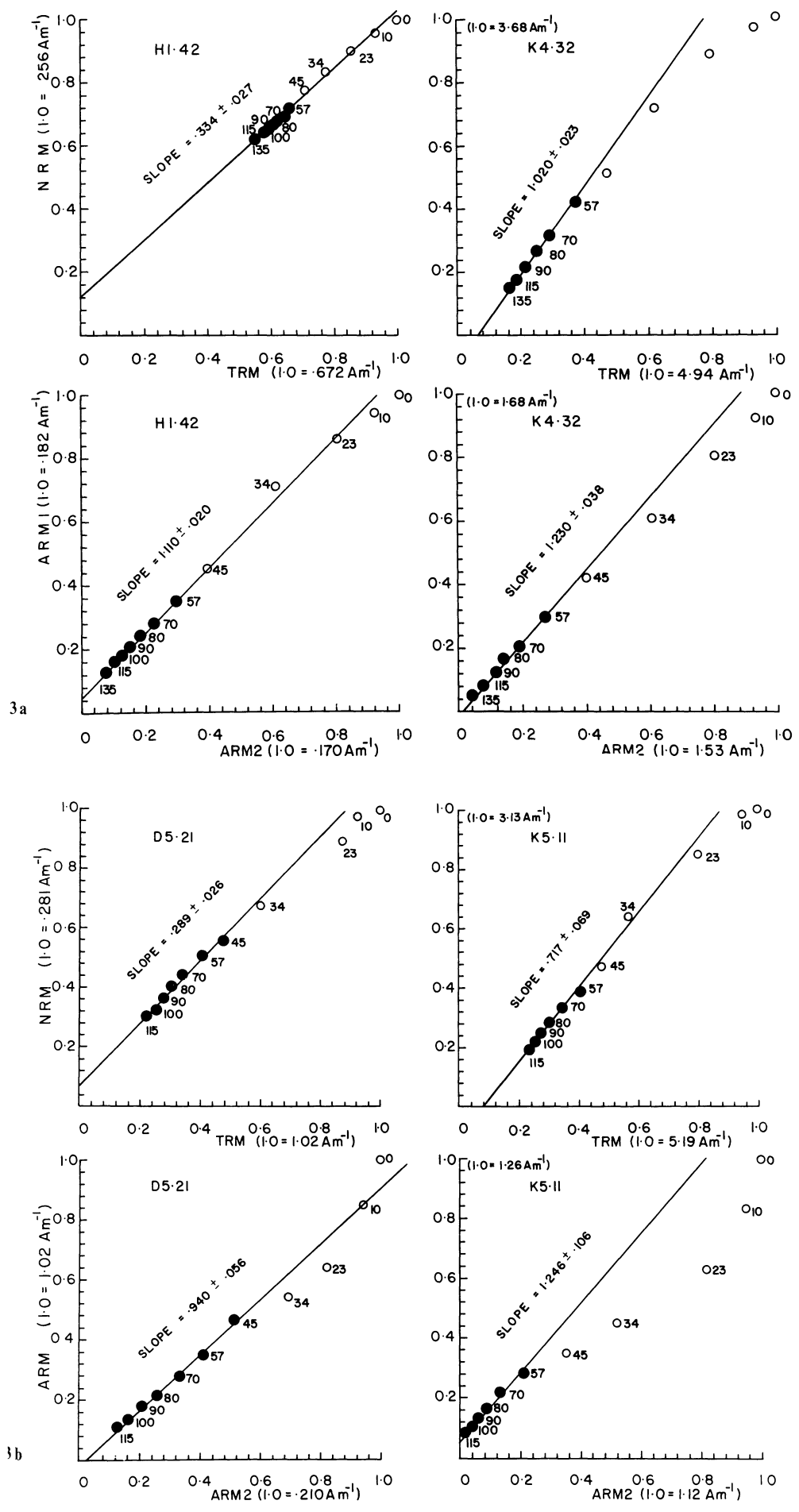
The procedure outlined above was applied to 32 specimens from the Exeter traps and the 9 specimens from the Esterel volcanics. Preliminary investigations had al-





b

Fig. 2a and b. Comparison of A.F. demagnetization curves of NRM and TRM for samples of a Exeter traps, b Esterel volcanics



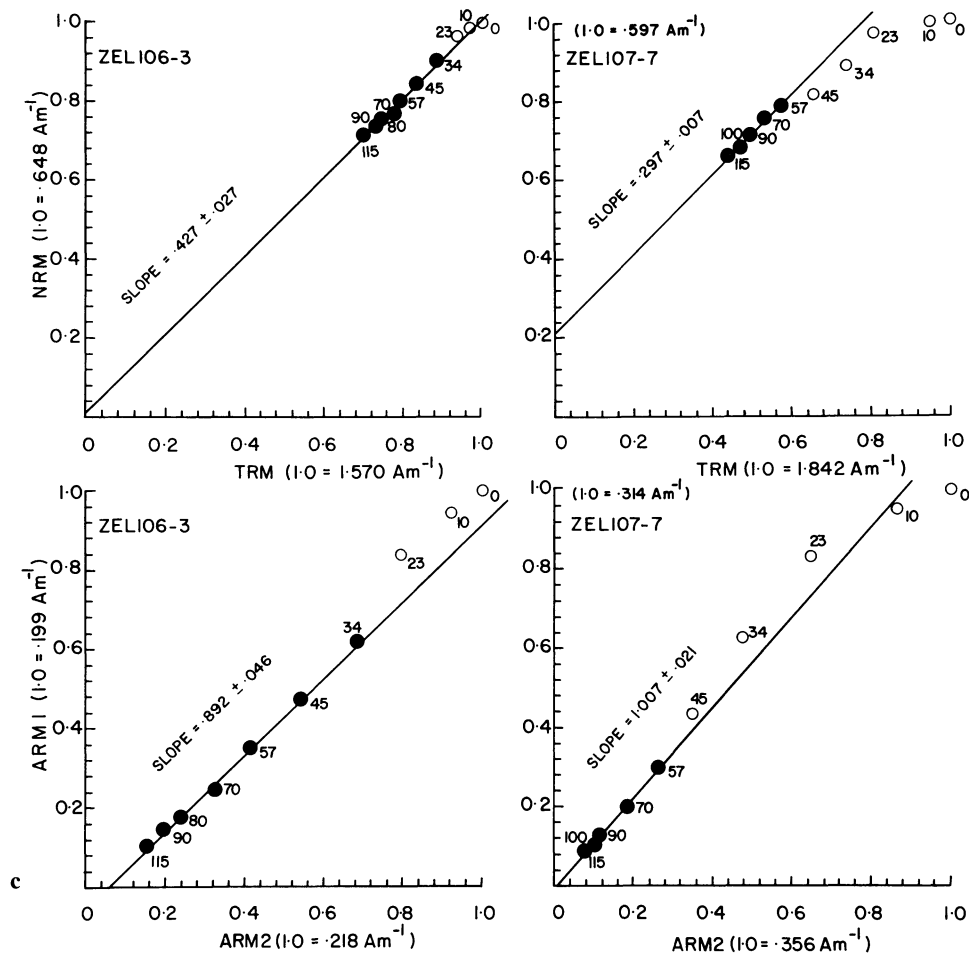


Fig. 3a-c. NRM-TRM and ARM1-ARM2 plots at various alternating fields (indicated in mT at each point) for **a** and **b** samples of the Exeter traps, and **c** samples of the Esterel volcanics. Solid points are those used to determine the slope of the line

ready been made of 43 specimens from the Exeter traps using peak alternating fields up to 200 mT (the previous limit available at A.N.U.). These results confirmed the viability of the technique and the complete procedure using peak fields up to 550 mT was then applied to a further 32 specimens the results from which are reported here.

Figure 2 shows normalized A.F. demagnetization. Curves of NRM and TRM for samples from each of the sites collected (Fig. 2a, Exeter traps; Fig. 2b, Esterel volcanics). The curves show the range of responses that are typical of these very oxidized basalts. In the Exeter traps, samples from localities D, H, and K show close correspondence between NRM and TRM whereas for locality P there is little agreement at the low coercivity end, the curves only beginning to merge above about 300 mT. On the other hand the Esterel volcanics show intermediate type responses to the demagnetization of NRM and TRM. Generally the TRM demagnetization is much steeper at low fields than the NRM.

Figure 3 compares plots of NRM versus TRM with ARM1 versus ARM2. The least squares slope in each case is measured over that range of alternating fields where there is a linear segment at higher fields. Standard errors are given for the slopes in each case. In virtually all cases the slopes of ARM1 versus ARM2 are close to the values 1.0 required for an ideal Shaw

palaeointensity, but in most cases they can be shown to be significantly different from 1.0 at the 95 percent confidence level. Thus some correction has to be made following Eq. (8). As Kono (1978) points out this assumes that the change in the coercivity spectrum of NRM has followed that of ARM.

Figure 4 now compares the magnetization curves of IRM1 versus IRM2 up to the maximum D.C. field available of 2.2 T. Generally speaking the data at high magnetizing fields greater than 0.5 T showed linear relationships. However in all cases the slopes in this region were significantly less than one, the only exception being results from sample P4 (Table 1). This indicates that there has been an increase in the amount of haematite present after heating in virtually every case. The slope of the IRM1-IRM2 plot between 0.5 and 2.2 T provides the appropriate correction factor to take account of this as given in Eq. (12). Because the IRM plots suggest that only haematite is involved at fields greater than about 500 mT, we have used the maximum peak alternating field of 550 mT only to determine the NRM/TRM ratio for the haematite high coercivity component. This involves the ratio of two measurements on the magnetometer, each of which is measured with about 5% accuracy. Therefore an accuracy of 7% is assumed for this ratio.

Tables 1 and 2 list the overall results for the two

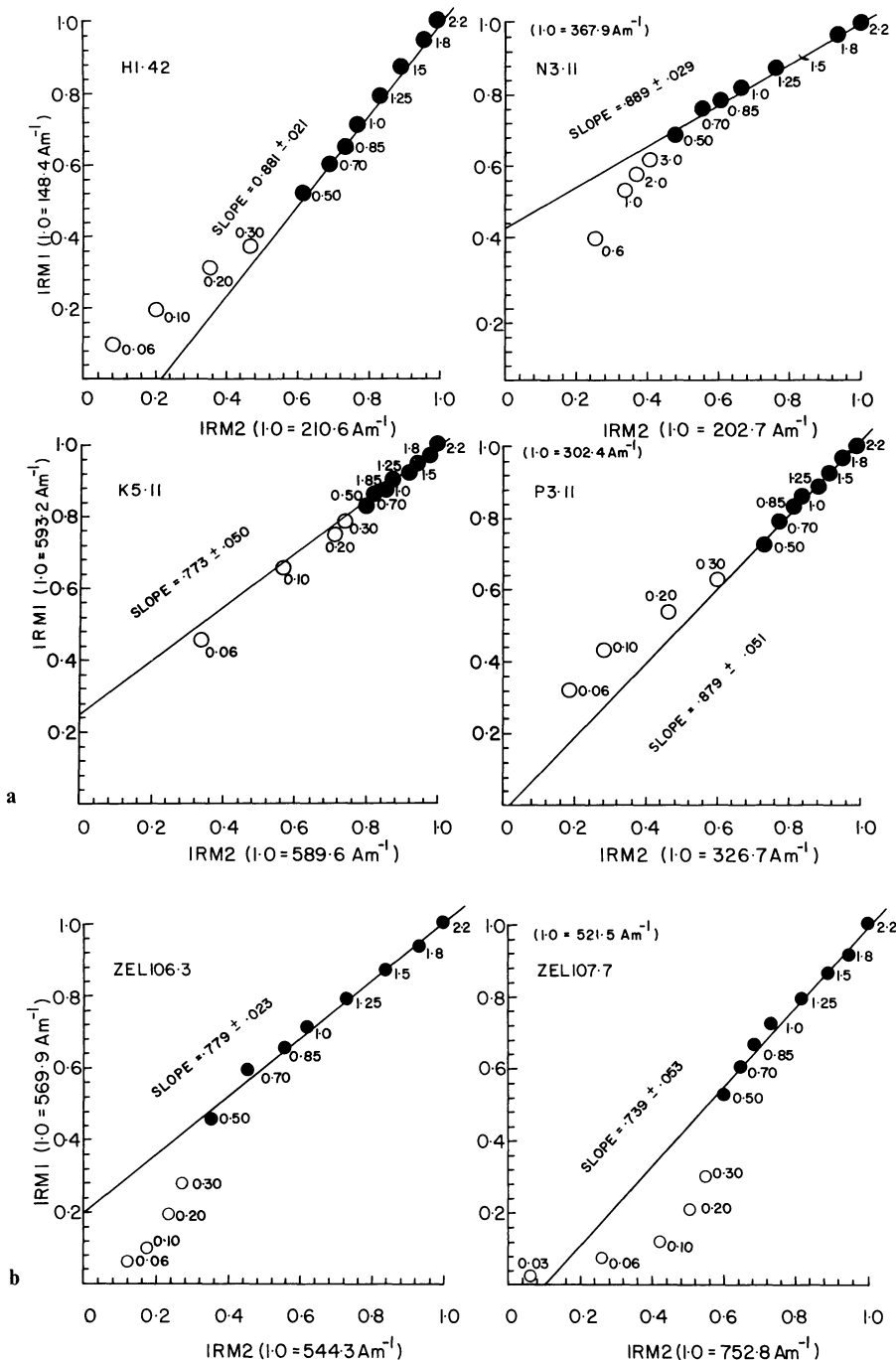


Fig. 4a and b. IRM1-IRM2 plots at various D.C. fields (indicated in T at each point) for samples of **a** Exeter traps, and **b** Esterel volcanics. The solid points are those used to determine the slope of the line

formations. The errors quoted are always standard errors and are similar for both the low and high coercivity determinations of palaeointensity, averaging about 10% (Fig. 5). The results for the two parts of the coercivity spectrum are not systematically higher or lower by either method, nor are the two values ever significantly different at the 95% confidence level. This encourages us to believe the method is a viable one, enabling the extension of Shaw's method to highly oxidized rocks. The results also provide further indication that the ARM correction method of Kono (1978) is a viable one. However they should not be taken to suggest that the IRM acquisition correction method by itself is always necessarily a viable one, and can be used in isolation.

Permian Dipole Moment

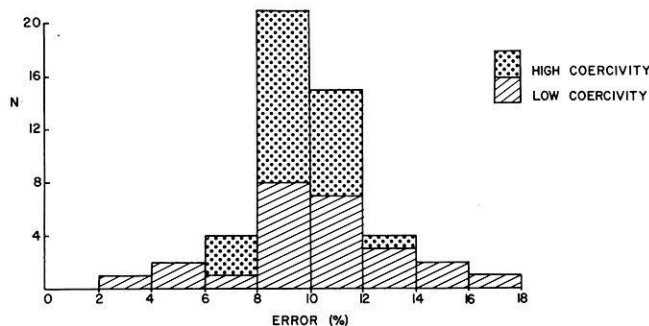
For each of the samples listed in Tables 1 and 2 a Virtual Dipole Moment (VDM) has been calculated using the observed values of inclination at each locality given by Zijdeveld (1967, 1975). The conversion of palaeointensity to VDM follows the proposal of Smith (1967a), who pointed out that this enabled the effects of dipole wobble to be eliminated when an average VDM is calculated. It is not clear at each of the localities in the Exeter traps whether or not more than one flow has been sampled. Certainly, the variation within each locality suggests this to be the case. Therefore in the Exeter traps all 21 values of VDM have been averaged to give a mean value of $6.11 \pm 1.12 \times 10^{22} \text{ Am}^2$ where the

Table 1. Palaeointensity values from the Exeter trap basalt samples. (F_1 , P_1 = palaeointensity and VDM from low coercivity region; F_2 , P_2 = palaeointensity and VDM from high coercivity region; VDM = Virtual Dipole Moment). Errors quoted are standard errors

Sample Number	Low Coercivity				High Coercivity				
	Slope of ARM1-ARM2	Slope of NRM-TRM	F_1 (μT)	P_1 (10^{22} Am 2)	Slope of IRM1-IRM2	$\left(\frac{\text{NRM}}{\text{TRM}}\right)_{550}$	F_2 (μT)	P_2 (10^{22} Am 2)	VDM (10^{22} Am 2)
	K1.41	1.289 ± 0.060	0.612 ± 0.081	28.15 ± 3.95	6.77 ± 0.95	0.742 ± 0.051	0.375	29.97 ± 2.94	7.21 ± 0.71
K2.31	1.224 ± 0.109	0.556 ± 0.047	26.95 ± 3.30	6.48 ± 0.80	0.739 ± 0.047	0.367	29.45 ± 2.79	7.09 ± 0.67	6.79
K3.12	1.181 ± 0.076	0.581 ± 0.034	29.19 ± 2.54	7.03 ± 0.61	0.751 ± 0.046	0.448	35.37 ± 3.29	8.51 ± 0.80	7.77
K4.32	1.230 ± 0.038	1.020 ± 0.024	49.19 ± 1.92	11.84 ± 0.46	0.698 ± 0.055	0.542	46.06 ± 4.81	11.09 ± 1.54	11.47
K4.43	1.219 ± 0.043	0.976 ± 0.031	47.49 ± 2.27	11.42 ± 0.55	0.722 ± 0.062	0.574	47.14 ± 5.12	11.34 ± 1.23	11.38
K5.11	1.246 ± 0.106	0.717 ± 0.069	34.13 ± 4.39	8.21 ± 1.06	0.773 ± 0.050	0.389	29.84 ± 2.81	7.18 ± 0.67	7.70
K5.72	1.198 ± 0.082	0.753 ± 0.052	37.27 ± 3.62	8.97 ± 0.87	0.741 ± 0.041	0.398	31.85 ± 2.84	7.66 ± 0.068	8.01
K6.12	1.168 ± 0.113	0.726 ± 0.052	36.86 ± 4.43	8.87 ± 1.07	0.695 ± 0.037	0.416	35.49 ± 3.12	8.54 ± 0.75	8.71
K6.31	1.189 ± 0.086	0.761 ± 0.061	37.96 ± 4.09	9.31 ± 0.98	0.670 ± 0.047	0.437	38.68 ± 3.83	9.31 ± 0.92	8.97
D1.13	0.891 ± 0.060	0.403 ± 0.019	26.82 ± 2.20	6.45 ± 0.53	0.739 ± 0.048	0.328	26.35 ± 2.52	6.34 ± 0.60	6.40
D1.31	0.914 ± 0.056	0.425 ± 0.028	27.57 ± 2.48	6.63 ± 0.60	0.771 ± 0.052	0.334	25.68 ± 2.50	6.18 ± 0.60	6.41
D3.41	0.828 ± 0.073	0.278 ± 0.018	19.91 ± 2.16	4.79 ± 0.52	0.716 ± 0.067	0.196	16.23 ± 1.90	3.90 ± 0.46	4.35
D4.13	0.880 ± 0.053	0.213 ± 0.025	14.35 ± 1.88	3.45 ± 0.45	0.685 ± 0.037	0.216	18.70 ± 1.65	4.50 ± 0.40	3.98
D5.21	0.940 ± 0.056	0.289 ± 0.026	18.25 ± 1.97	4.39 ± 0.47	0.703 ± 0.058	0.230	19.40 ± 2.09	4.67 ± 0.50	4.53
D5.41	0.932 ± 0.043	0.272 ± 0.028	17.31 ± 1.95	4.16 ± 0.46	0.732 ± 0.060	0.245	19.84 ± 2.14	4.77 ± 0.51	4.47
H1.42	1.110 ± 0.020	0.334 ± 0.027	17.84 ± 1.48	4.29 ± 0.36	0.881 ± 0.021	0.285	19.28 ± 1.42	4.64 ± 0.34	4.47
H2.51	1.083 ± 0.051	0.382 ± 0.024	20.92 ± 1.64	5.03 ± 0.40	0.860 ± 0.041	0.295	20.34 ± 1.72	4.89 ± 0.41	4.96
H2.22	1.067 ± 0.069	0.356 ± 0.031	19.80 ± 2.15	4.76 ± 0.52	0.881 ± 0.039	0.292	19.65 ± 1.63	4.73 ± 0.39	4.75
H3.41					0.831 ± 0.065	0.278	19.83 ± 2.08	4.77 ± 0.50	4.77
H4.51					0.998 ± 0.044	0.305	18.19 ± 1.50	4.38 ± 0.36	4.38
H4.12					0.978 ± 0.041	0.308	18.68 ± 1.53	4.50 ± 0.37	4.50
N1.11					0.898 ± 0.046	0.195	12.87 ± 1.11	3.10 ± 0.27	3.10
N3.11					0.889 ± 0.030	0.162	10.81 ± 0.84	2.61 ± 0.20	2.61
N3.21					0.872 ± 0.041	0.140	9.52 ± 0.80	2.29 ± 0.19	2.29
N4.42					0.891 ± 0.052	0.106	7.06 ± 0.64	1.71 ± 0.15	1.71
P1.21					0.792 ± 0.038	0.330	24.70 ± 2.10	5.94 ± 0.51	5.94
P1.11					0.801 ± 0.045	0.324	23.99 ± 2.15	5.77 ± 0.52	5.77
P2.21					0.828 ± 0.061	0.453	32.41 ± 3.29	7.80 ± 0.79	7.80
P3.11					0.879 ± 0.051	0.572	38.58 ± 3.51	9.28 ± 0.84	9.28
P3.21					0.858 ± 0.056	0.535	36.98 ± 3.54	8.90 ± 0.85	8.90
P4.22					0.974 ± 0.066	0.738	44.93 ± 4.38	10.81 ± 1.05	10.81
P4.41					0.982 ± 0.059	0.721	43.54 ± 4.02	10.48 ± 0.96	10.48

Table 2. Palaeointensity values from the Esterel volcanics. (I* = palaeoinclination obtained from Zijdeveld, 1975). The other symbols are as in Table 1.

Sample Number	I*	Low Coercivity				High Coercivity				F_2 (μ T)	P_2 (10^{22} Am ²)	VDM (10^{22} Am ²)
		Slope of ARM1-ARM2	Slope of NRM-TRM	F_1 (μ T)	P_1 (10^{22} Am ²)	Slope of IRM1-IRM2	$\left(\frac{\text{NRM}}{\text{TRM}}\right)_{550}$					
ZEL 106-3	-29.5°	0.892 ± 0.046	0.427 ± 0.027	28.39 ± 2.31	6.64 ± 0.54	0.779 ± 0.023	0.324	24.66 ± 1.87	5.77 ± 0.44	6.21		
ZEL 106-4	-29.5°	0.907 ± 0.065	0.473 ± 0.036	30.92 ± 3.23	7.23 ± 0.76	0.789 ± 0.047	0.311	23.10 ± 2.12	5.40 ± 0.50	6.32		
ZEL 107-7	-29.5°	1.007 ± 0.020	0.297 ± 0.006	17.49 ± 0.50	4.09 ± 0.12	0.739 ± 0.053	0.232	18.59 ± 1.86	4.35 ± 0.44	4.22		
ZEL 107-11	-29.5°	0.978 ± 0.53	0.316 ± 0.019	19.16 ± 1.55	4.49 ± 0.36	0.758 ± 0.070	0.258	20.18 ± 2.34	4.72 ± 0.55	4.61		
ZEL 26-9	-19.5°	0.939 ± 0.019	0.106 ± 0.003	6.70 ± 0.24	1.68 ± 0.07	-	-	-	-	1.68		
ZEL 89-6	-25.0°	1.060 ± 0.091	0.235 ± 0.039	13.15 ± 2.44	3.17 ± 0.59	0.642 ± 0.053	0.210	19.40 ± 2.10	4.68 ± 0.51	3.93		
ZEL 97-2	-21.5°	1.120 ± 0.077	0.252 ± 0.028	13.34 ± 1.74	3.27 ± 0.43	0.812 ± 0.053	0.257	18.76 ± 1.80	4.61 ± 0.44	3.95		
ZEL 97-4	-21.5°	1.176 ± 0.061	0.268 ± 0.037	13.51 ± 1.99	3.32 ± 0.49	0.829 ± 0.041	0.241	17.24 ± 1.48	4.23 ± 0.36	3.78		
ZEL 22-5	-	Non-Linear	Non-Linear	-	-	-	-	-	-	-		

**Fig. 5.** Histogram of the standard errors for each palaeointensity determination using the high coercivity or low coercivity fraction

error quoted is the 95% confidence limit. The standard deviation is 2.62×10^{22} Am² which is 42.9% of the mean. This is typical of the scatter observed in VDMs over the last 5 million years (McFadden and McElhinny, 1982) and again suggests that fluctuations in dipole intensity are the main source of this scatter.

In the Esterel volcanics the five sites yield a mean VDM of $4.03 \pm 1.43 \times 10^{22}$ Am² with error given at the 95% confidence level. The standard deviation is 1.64×10^{22} Am² which is 40.6% of the mean. The mean VDM is significantly different from that for the Exeter traps, but this may not be very significant because archaeomagnetic data suggest wide variations in dipole moment over times of $<10^5$ years (McElhinny and Senanayake, 1982). The data from these two formations is probably inadequate to provide a good mean dipole moment for the Permian. However, both values are somewhat higher than those predicted from previous palaeointensity data for the Phanerozoic (Smith, 1967b; McElhinny, 1973).

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