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The magnitude of the palaeomagnetic field in Iceland between 2 and 6 Myr ago

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Summary. Sixty-eight palaeomagnetic field magnitude values have been determined from a sequence of Icelandic lavas, ranging from 2 to 6 Myr in age. The results indicate large and rapid changes in the palaeomagnetic dipole field and provide a mean value of the palaeomagnetic field magnitude in Iceland for this period.

Introduction

Studies of polarity transitions (Shaw 1975, 1977; Shaw & Wilson 1977) have shown that the magnitude of the field can vary greatly, even though the direction maintains a constant intermediate value. This study applied the same technique for determining the magnitude of the palaeomagnetic field (Shaw 1974) to see if large changes in magnitude occurred during the normal and reversed states.

An extensive collection of cores from some 900 individual lavas extruded during the period 2–13 Myr ago had already been made for palaeomagnetic direction purposes (Dagley *et al.* 1967). This work determined the polarity and time order of the lavas and therefore most of the collected cores had been subjected to alternative field (af) demagnetization to peak fields of about 0.04 T. The technique used to determine the magnitude of the palaeomagnetic field (palaeofield) also relies on af demagnetization (Shaw 1974) but to much higher values, up to 0.25 T, so the existing cores were quite suitable for this further study.

One-hundred and ninety-nine cores from the seven youngest sections were used for this study (Fig. 1). The sections have been dated by the $^{40}Ar/^{39}Ar$ method (Mussett, Ross & Gibson 1980; Ross & Mussett 1976) and range from 2 to 6 Myr in age.

Field magnitude measurements

The technique used to determine the palaeofield magnitude (Shaw 1974, 1979) employs progressive af demagnetization and measurement of both the NRM and a full laboratory-induced TRM to enable comparison between equivalent regions of the coercive force spectrum. Any thermal alteration resulting from the laboratory heating used to give the

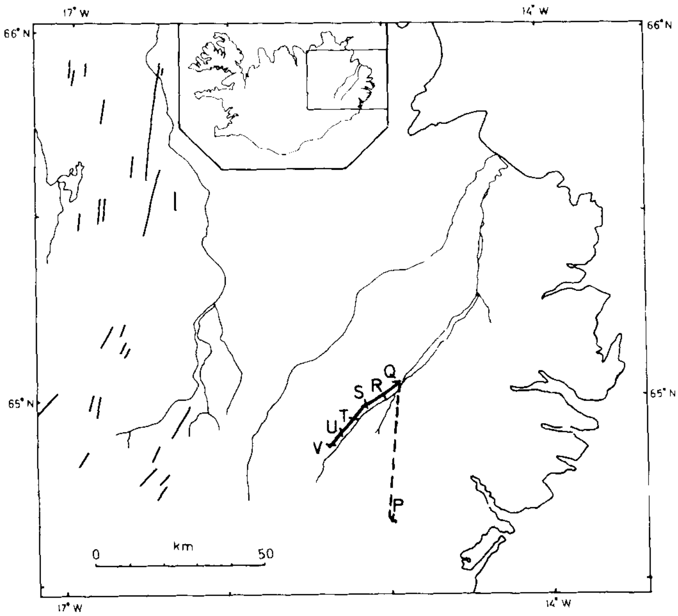

Figure 1. Map of eastern Iceland. The large letters show the locations of some of the sections sampled by Dagley *et al.* (1967), from whose collection samples were taken for both the dating (Mussett, Ross & Gibson 1980; Ross & Mussett 1976) and for the field magnitude analysis. The dotted line indicates the sequential link between sections P and Q.

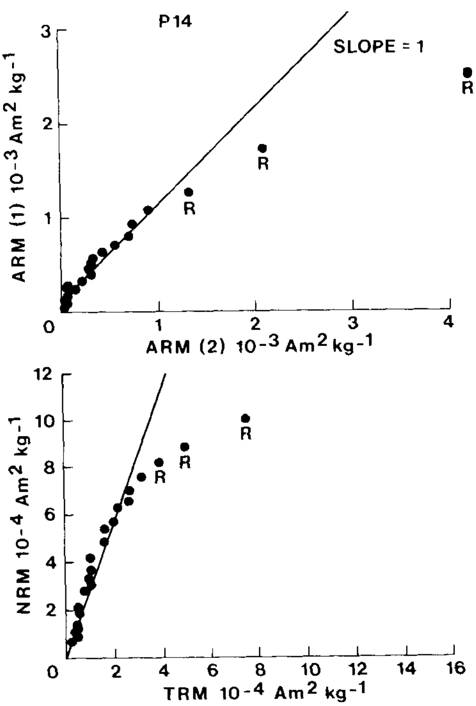

Figure 2. All of the results from sample P14. The upper graph of ARM (1) against ARM (2) detects thermal alteration in the low coercive force region (marked R for rejected). The lower graph is a plot of NRM against TRM; the altered coercive force region detected by the ARMs is rejected (also marked R) and the best fitting line calculated from the remaining points and the origin.

TRM can be detected by comparing two anhysteretic remanent magnetizations (ARMs) given to the same specimen. ARM (1) is given before heating and after af demagnetization of the NRM. ARM (2) is given after heating and after af demagnetization of the TRM.

By choosing a region of the ARM (1)/ARM (2) graph which has slope = 1 (i.e. no thermal alteration in that region) a comparison of NRM/TRM *in that unaltered region* will allow the ancient field magnitude to be determined, using equation (1), which is valid for magnetic fields up to 100 μT (Nagata 1943), i.e. to about twice the strength of the present field. A typical example of this work is plotted in Fig. 2.

$$\frac{\mathrm{TRM}}{\mathrm{NRM}} = \frac{\text{laboratory field magnitude}}{\text{palaeomagnetic field magnitude}}. \tag{1}$$

The ARMs are not used to determine the palaeofield magnitude directly; they are simply used to detect thermal alteration of the coercive force spectrum.

Of the 199 cores measured, 43 cores had NRMs which were not stable when subjected to high af demagnetization. Empirically, only samples with a measurable remanence after af demagnetization in 0.1 T are suitable for palaeofield magnitude studies; measurements on the 43 unstable cores were not continued beyond the NRM measurement.

Of the remaining 156 cores, eight exploded beyond recovery and 80 became thermally altered throughout the whole of their observable coercive force spectrum. The remaining 68 cores satisfied the above criteria, but three (U12, S33 and R1A) of the magnitude values were very much larger than 100 μT. Since Nagata's (1943) results indicate that the proportionality implicit in equation (1) may break down above 100 μT these three samples were also given TRM's in high fields (Table 1), and the ARM (2)s were repeated to check for thermal alteration.

For sample U12 the laboratory fields straddled the deduced ancient fields (Table 1) and as the two values of the ancient field were similar it seems that non-proportionality was small and the mean of 130.4 μT, was used. For S33 both laboratory fields were less than the deduced fields which differed by about 6 per cent. Because of the lack of proportionality the values are likely to be low, and the larger of the two, 154.8 μT, was adopted. The difference of the two values for R1A was so marked that the departure from proportionality must have been large and perhaps not simple, and the larger of the two values, 220.8 μT, is adopted with reservations. These three samples yield peaks of field magnitude (Fig. 3); the reality of the peak at U12 is supported by adjacent samples but those of R1A and S33 stand alone.

Apart from these three samples, the laboratory fields used to determine magnitudes were either 38 or 50 μT.

Table 1. The samples that were re-heated in stronger laboratory fields than usual because they recorded palaeofield magnitude values much greater than 100 μT.

Sample	Laboratory magnitude (μT)	Ancient field magnitude (μT)	Standard deviation (μT)	Comment
U12	50.0	131.9	5.5	Mean value of
U12	200.0	128.8	4.0	130.4 μT used
S33	50.0	124.0	7.3	High value of
S33	115.0	154.8	2.8	154.8 μT used
R1A	38.0	147.2	21.3	High value of
R1A	115.0	220.8	9.9	220.8 μT used

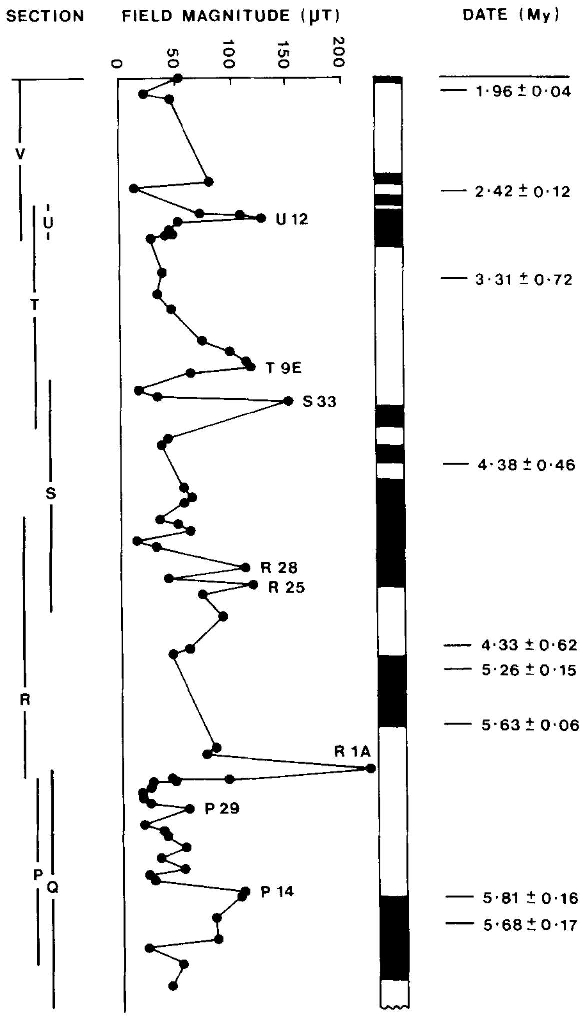

Figure 3. Composite diagrams with the time axis based on the assumption that on average the lavas are extruded at regular intervals. The $^{40}Ar/^{39}Ar$ dates confirm this assumption. The time overlap of the sampling sections are shown together with the individual palaeofield magnitude data and the polarity of all measured samples (dark areas represent normal polarity).

Results and analysis

The 68 acceptable field magnitude results are listed in chronological order (Dagley *et al.* 1967) in Table 2. The standard deviations were obtained directly from the straight line fit of the NRM/TRM data. The results are plotted in Fig. 3, together with polarities and heights of the sections made proportional to their numbers of lavas. The time-scale assumes a linear rate of extrusion of lavas and is based upon the $^{40}Ar/^{39}Ar$ dates of Mussett *et al.* (1980). As these dates have errors up to 0.72 Ma and the rate of extrusion of lavas probably was irregular (Mussett *et al.* 1980), durations of intervals given are approximate.

It is clear that there have been large changes in palaeofield magnitude, covering about an order of magnitude. The variation does not follow a normal distribution (Figs 3 and 4) and there is a tendency for positive swings of the magnitude to be both larger and sharper than negative swings.

Table 2. The individual palaeofield magnitude values, standard deviation and polarity together with the mean (M), standard deviation (σ) and standard error on the mean (ϵ) for the four long polarity intervals.

Code	Field Magnitude (μT)	Standard Deviation (μT)	Polarity	M (μT)	σ (μT)	ϵ (μT)
V19	55.4*	4.5	N			
V16	23.5*	0.8	R			
V15	46.9*	1.6	R			
V10	82.7*	1.3	N			
V 9	14.1*	1.0	N			
T29	72.5	3.4	N			
U13	110.0	20.2	N			
U12	130.4**	5.0	N	66.5	34.6	11.4
U 9	54.9	2.1	N			
V 4	44.0*	0.9	N			
U 4	47.8	2.1	N			
U 2	42.3	1.2	N			
V 2	27.4*	14.0	R			
T18	37.0	3.9	R			
T15	33.5	1.0	R			
T13A	45.4	2.4	R			
T 9J	74.6	3.0	R			
T 9H	99.7	6.3	R			
T 9F	114.3	12.6	R			
T 9E	117.4	11.9	R	63.5	40.7	10.9
T 9D	62.8	3.3	R			
S37	14.0*	0.4	R			
S36	31.4*	3.5	R			
S33	154.8**	2.8	R			
S25	41.8*	2.4	R			
S24	35.5*	2.7	R			
S17D	56.0*	2.9	N			
S17B	64.7*	3.1	N			
S17A	56.2*	1.6	N			
R33D	34.2	1.8	N			
R33C	50.5	1.8	N			
R33B	61.5	3.3	N	58.3	31.3	9.4
R33	12.6	0.4	N			
R32	30.3	1.2	N			
R28	115.9	22.5	N			
R26	41.7	0.7	N			
R25	118.1	16.9	N			
R23	72.0	4.4	R			
R20	90.3	4.8	R			
R13	60.0	1.8	R			
R12	44.7	1.2	R			
R 1F	84.0	14.2	R			
R 1E	75.4	1.6	R			
R 1A	220.8**	9.9	R			
R 1	93.4	15.0	R			
P35	45.1	0.7	R			
P34	48.1	1.9	R			
Q11	25.8	2.3	R			
Q10	22.3	14.7	R			
P32	15.6	0.8	R			
P31	15.2	0.8	R			
P30	23.1	3.7	R			
P29	58.0	1.3	R			
P26	18.0	1.1	R			
P25	35.4	1.4	R	57.3	42.8	8.4
P24	38.3	0.8	R			
P22	55.8	0.6	R			
P20	31.4	0.7	R			
P18	53.4	0.8	R			
P17	22.6	7.1	R			
P16	26.6	3.6	R			
P14	108.2	9.7	R			
P13	104.6	5.1	R			
P 9	82.1	5.1	N			
P 6	82.9	7.3	N			
P 5	20.0	3.1	N			
P 1	52.8	2.2	N			
Q 4	42.2	1.1	R			

* Indicates a laboratory field of 50 μT.

** See Table 1 for laboratory field.

The average field magnitude over the 4 Ma period is 59 μT, with a standard deviation, σ, of 38 μT and a standard error of the mean, ϵ, of 5 μT. Assuming an axial geocentric dipole, this is equivalent to a dipole moment of 7.7×10^{22} A m^2, which is considerably larger than Smith's (1967a) estimate for the Upper Tertiary virtual dipole moment of 5.53×10^{22} A m^2.

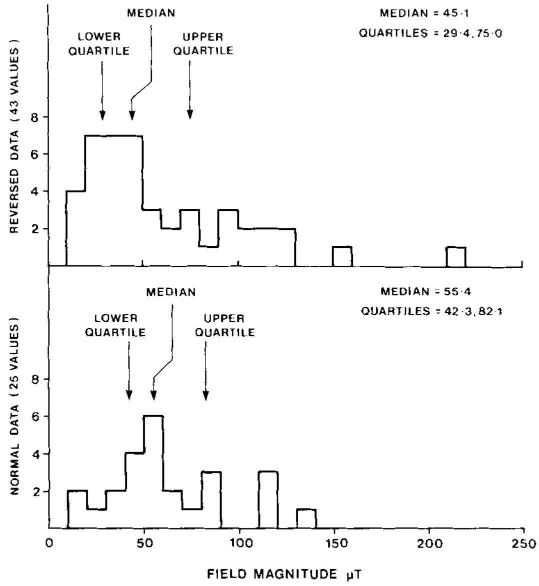

Figure 4. Histograms, showing the frequency of occurrence of field magnitude values in both the normal and reversed states. These graphs are, in effect, probability distributions of particular field values occurring. The data suggest that low field values are more likely to occur during reversed periods.

The normal field magnitudes (mean = 61 μT, σ = 31 μT, ϵ = 6 μT) do not appear to be significantly different from those of the reversed field (mean = 56 μT, σ = 41 μT and ϵ = 6 μT), in agreement with earlier work (Smith 1967b). However, these statistics disguise the fact that the distribution is not Gaussian, and histograms (Fig. 4) show considerable differences between normal and reversed magnitudes, the mode, median and the quartiles of the reversed field values being lower than those of the normal field values.

Four of the polarity intervals each contain sufficient data for the mean, standard deviation and standard error to be calculated (Table 2). The reversed interval means (63.5 and 57.3 μT) are on average lower than the normal means (66.5 and 58.3 μT).

Although the four polarity interval means suggest a decrease in field magnitude with age, when the data are divided into two (youngest 34 and oldest 34 data) their means are almost the same (58.2 and 60.1 μT respectively).

Because lava extrusion is intermittent the magnitudes are in the nature of spot values, with unknown values between. Archaeomagnetic measurements (e.g. Games 1980) show that the field can double in 200 yr, an interval beyond the resolution of the lava piles. However, during the two sequences of lavas T9E to U12 and P29 to R1A, estimated to be 70 000 and 20 000 yr long respectively, the magnitude changes smoothly, suggesting shorter term variations cannot be large. On the other hand, in other parts of the sequence, such as P17 to P29 and R20 to R32 the full variation clearly is not being recorded.

Discussions and conclusions

The 68 results presented in this paper have established a mean value for the palaeomagnetic field magnitude in Iceland during the period 2–6 Myr. This mean value corresponds to an average axial dipole moment of 7.7×10^{22} A m^2 which is significantly larger than Smith's (1967a) estimate of the virtual dipole moment (5.35×10^{22} A m^2).

The separate averages of the normal and reversed data are not significantly different but there is a suggestion of some asymmetry in the normal and reversed field states in that the mode, median and quartiles of the reversed data are lower than those of the normal data. Watkins & Haggerty (1978) observed a strong correlation between high oxidation states and reversed lavas in Eastern Iceland. The suspected asymmetry of this palaoefield magnitude reported here may be connected in some way with this observation.

Finally, we have observed large changes in the palaeofield magnitude during normal and reversed polarity intervals and these do not have a Gaussian distribution. These changes are somewhat larger than those observed over archaeological time (see, for example, Games 1980 and Shaw 1979) as might be expected. The range of periodicity of these field magnitude fluctuations is difficult to determine accurately because of the intermittent nature of the data and the errors on the dates. The longest periods of smooth change are between approximately 2 and 7×10^4 yr but the majority of fluctuations have much shorter periods than this.

Cande & Labreque (1974) observed sequences of short wavelength, low amplitude seafloor magnetic anomaly patterns of constant polarity. Because of the global distribution of these anomalies they deduced that they were due to time variations of the geomagnetic dipole moment. They were able to model the observed anomalies using geomagnetic field amplitude variations equal or greater than 15 per cent with a periodicity equal or greater than 3×10^4 yr. They also noted that the observed anomaly pattern could equally well be modelled by shorter period fluctuations of greater amplitude. It is possible therefore that the short period (less than 10^4 yr) large amplitude (up to 200 per cent of the mean) variations together with the longer period smooth variations of Fig. 3 are similar to the field magnitude variations that produced the short wavelength anomaly pattern observed by Cande & Labreque. If this is the case then the global distribution of the seafloor data would lead us to believe that our observations of changes in the field magnitude are due primarily to changes in the dipole field, not to local changes.

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References

Cande, S. C. & Labreque, J. L., 1974. Behaviour of the Earth's palaeomagnetic field from small scale marine magnetic anomalies, *Nature*, **247**, 26–28.

Dagley, P., Wilson, R. L., Ade-Hall, J. M., Walker, G. P. L., Haggerty, S. E., Sigurgeirsson, T., Watkins, N. D., Smith, P. J., Edwards, J. & Grasty, R. L., 1967. Geomagnetic polarity zones for Icelandic lavas, *Nature*, **216**, 25–29.

Games, K. P., 1980. The magnitude of the archaeomagnetic field in Egypt between 3000 and 0 BC, *Geophys. J. R. astr. Soc.*, **63**, 45–56.

Mussett, A. E., Ross. J. G. & Gibson, I. L., 1980. $^{40}Ar/^{39}Ar$ dates of eastern Iceland lavas, *Geophys. J. R. astr. Soc.*, **60**, 1–19.

Nagata, T., 1943. The natural remanent magnetism of volcanic rocks and its relation to geomagnetic phenomena, *Bull. earthq. Res., Inst. Tokyo*, **21**, 1–196.

Ross, J. G. & Mussett, A. E., 1976. $^{40}Ar/^{39}Ar$ dates for spreading rates in eastern Iceland, *Nature*, **259**, 36–38.

Shaw, J., 1974. A new method of determining the magnitude of the palaeomagnetic field: application to 5 historic lavas and 5 archaeological specimens, *Geophys. J. R. astr. Soc.*, **39**, 133–141.

Shaw, J., 1975. Strong geomagnetic fields during a single Icelandic polarity transition, *Geophys. J. R. astr. Soc.*, **40**, 345–350.

Shaw, J., 1977. Further evidence for a strong intermediate state of the palaeomagnetic field, *Geophys. J. R. astr. Soc.*, **48**, 263–269.

Shaw, J., 1979. Rapid changes in the magnitude of the archaeomagnetic field, *Geophys. J. R. astr. Soc.*, **58**, 107–116.

Shaw, J. & Wilson, R. L., 1977. The magnitude of the palaeomagnetic field during a polarity transition: a new technique and its application, *Phys. Earth planet. Int.*, **13**, 339–341.

Smith, P. J., 1967a. The intensity of the tertiary geomagnetic field, *Geophys. J. R. astr. Soc.*, **12**, 239–258.

Smith, P. J., 1967b. The intensity of the ancient geomagnetic field: a review and analysis, *Geophys. J. R. astr. Soc.*, **12**, 321–362.

Watkins, N. D. & Haggerty, S. E., 1968. Oxidation and magnetic polarity in single Icelandic lavas and dykes, *Geophys. J. R. astr. Soc.*, **15**, 305–316.